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J.E. CHRISTOPHER AND F.M. HAIDL

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FOREWORD

The "First International Williston Basin Symposium" was held in Regina thirty-five years ago. Little did the organizers of that event realize that they were starting a long-standing tradition which would foster a spirit of co-operation and collegiality across the international border. The results have been of far greater mutual benefit than any free trade agreement will ever be!

Despite the youthfulness of the Williston Basin "oil play" in 1956, some 28 papers were presented at that initial symposium. A review of the table of contents of the first proceedings volume shows a focus on field studies and on regional stratigraphy covering all parts of the sedimentary succession. Tables of contents of subsequent symposia volumes demonstrate that Mississippian rocks and oil fields have attracted most attention, and that other topics have generated recurring interest. At least one paper on salt solution, on Cambro-Ordovician stratigraphy, and on the Winnipegosis Formation have been presented at each symposium. Ideas on the origin of Winnipegosis buildups have evolved from the radical hypothesis that they were dolomitic relicts of salt metasomatism of carbonate banks to their representing former mud mounds and reefs. That these same topics are on the agenda for this conference should surprise no-one.

The interval between successive symposia increased significantly after the second, which took place in 1958. The third (the first in which the Montana Geological Society participated in a sponsoring role) was held in Regina in 1964, the fourth (again in Regina) in 1982, and the fifth in Grand Forks in 1987. These last two gatherings each added new dimensions to the program: the fourth included, for the first time, pre- and post-symposium field trips; the fifth, a pre-symposium field trip and a post-symposium core workshop. The innovation for this sixth symposium is poster displays which have been so enthusiastically accepted that some 20 presentations are offered.

The Williston Basin is now a mature oil province. It is generally considered to be the archetypal intracratonic basin. Despite the volumes of information about the basin which have been published, each symposium continues to draw high levels of interest and has made valuable contributions to our understanding of this intriguing geological feature. The success of each symposium has been a credit to its hard-working organizers, and it is these people to whom this volume is dedicated.

Don Kent
Chairman,
Sixth International Williston
Basin Symposium

EDITORIAL COMMENTS

The papers incorporated in this volume were submitted four months prior to their presentation at the Sixth International Williston Basin Symposium. They were edited, processed, printed and bound in time for distribution to delegates attending the opening session. Accordingly, many of the standard courtesies in the editorial process, such as peer review for content and feed-back to the authors were sacrificed due to the brevity between submission of papers and their publication. We hope that undue liberties were not taken with authors' intents or meanings. As in previous volumes, data quality and interpretations remain a prerogative of the responsible author(s).

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- Biennial symposia, generally with themes which focus on different aspects of Saskatchewan's economic geology
- Special publications, mostly symposia proceedings, but including field excursion guides, geological highway maps (an upcoming project) and other miscellaneous studies
- Field trips
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- Talks to grade-school and high-school students, upon request
- Occasional public lectures
- Annual awards to the most promising earth sciences undergraduate at the Universities of Saskatchewan and Regina
- Annual Christmas lecture for Regina school students featuring a distinguished North American geoscientist
- Award for the best student poster display at the annual Saskatchewan Energy and Mines Open House

Notes

SUBSIDENCE CONTROLLED STRATIGRAPHIC SEQUENCES AND THE ORIGIN OF SHELF SAND RIDGES, WINNIPEG GROUP (MIDDLE ORDOVICIAN) MANITOBA, SASKATCHEWAN, NORTH DAKOTA

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Littleton, Colorado 80160-0269, USA

ABSTRACT

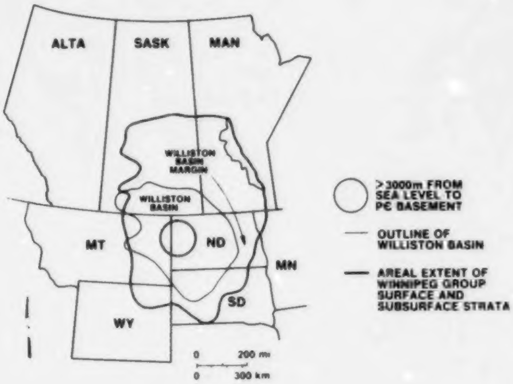
In southern Manitoba and eastern North Dakota, three large (kilometers in length and width) elongate east-west trending sand bodies of probable storm and tidal-current origin have been deposited on mostly muddy continental shelf deposits of the Icebox Formation of the Middle Ordovician (Caradoc) Winnipeg Group in the Williston Basin. Several smaller stratigraphically equivalent sand bodies of similar origin are observed in western Manitoba and southern Saskatchewan. These sand bodies all occur in the second of three subsidence controlled depositional sequences in the Winnipeg Group.

The largest of the sand bodies, the Carman sand in southern Manitoba, is approximately 30 m thick, 250 km long, and 45 km wide and was identified by using well log correlation and sample description. This sand body consists of at least five amalgamated sand ridges which accreted in a N to NNW direction. Individual sand ridges comprise a lower unit of bioturbated sandy siltstones (gradational with the underlying muddy shelf), grading upward into fine-grained, less silty sandstone of probable storm-current origin, and an upper unit of medium to coarse-grained, well-sorted, clean sandstone at the irregular top of the ridge. This clean sandstone was probably deposited by strong periodic or regular currents generated in a tidal regime. About 150 km to the south of the Carman sand in eastern North Dakota, two smaller elongate sand bodies in the Icebox Formation are of similar storm and tidal origin. These two ridges may be outer shelf equivalents of the Carman sand. The sand bodies in southern Saskatchewan are cross-bedded and burrowed and up to 5.0 m thick, 5.0 km long, and 2.0 km wide.

The three Winnipeg Group depositional sequences (W1-W3), which are loosely constrained by conodont biostratigraphy, are the result of episodes of basin margin subsidence and sediment input. Early Williston Basin thermally controlled subsidence and subsequent basin margin erosion were the principal sediment supply mechanisms. Shelf ridge development in the Icebox Formation was related to a lower than usual basin subsidence rate for the Middle Caradoc eastern Williston Basin margin during transgression. This slow subsidence resulted in the development of shoaling sand ridges to a critical depth for the deposition of tidal sediments.

INTRODUCTION

The purpose of this study is to define more specifically the depositional environment of the Icebox Formation of the Middle Ordovician (Caradoc) Winnipeg Group of the Williston Basin in Manitoba, Saskatchewan, and North Dakota (Figs. 1, 2). Three large (kilometers - 10's of kilometers in length and width) elongate east-west trending sand bodies in the Icebox Member in Manitoba and North Dakota are discussed in terms of new and evolving concepts of shelf sedimentation. Their relationship to early Williston Basin subsidence and history is placed in a sequence stratigraphic framework.


ALTA
SASK
MAN
WILLISTON BASIN MARGIN
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MT
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WY
>3000m FROM SEA LEVEL TO P€ BASEMENT
OUTLINE OF WILLISTON BASIN
AREAL EXTENT OF WINNIPEG GROUP SURFACE AND SUBSURFACE STRATA
0 200 mi
0 300 km

Figure 1. Williston Basin and areal extent of Winnipeg Group.

The Winnipeg Group is a sandstone-shale-marly limestone, up to 105 m thick, that unconformably overlies Precambrian metamorphics and Cambro-Ordovician sediments and underlies gradationally the Red River carbonates of Ashgill age.

PREVIOUS WORK

Because outcrop control is limited to the fringes of the Williston Basin in Manitoba, Saskatchewan, and South Dakota and well control within the basin (622 wells or one well per 1100 km^2) is sparse, detailed interpretations of the depositional environment within the Winnipeg Group are nonexistent. Paterson (1971), Vigrass (1971), McCabe (1978), and Kessler (1978) have made generally regional interpretations in Manitoba and Saskatchewan, as based on well log correlations and analysis of limited core data (only 10 cores in Manitoba and 41 in Saskatchewan penetrate any part of the Winnipeg). Vigrass (1971) offered the most extensive interpretation of the Winnipeg, suggesting that its deposition throughout was a period of overall marine transgression, with minor regressive phases occurring at the basin margin, especially during deposition of the lower Winnipeg or Black Island Formation. Sand was transported into the basin from the northern and eastern margins. He saw the upper Winnipeg or Icebox Formation, as onlapping the Black Island, but also with similar regressive sandstones along the periphery and

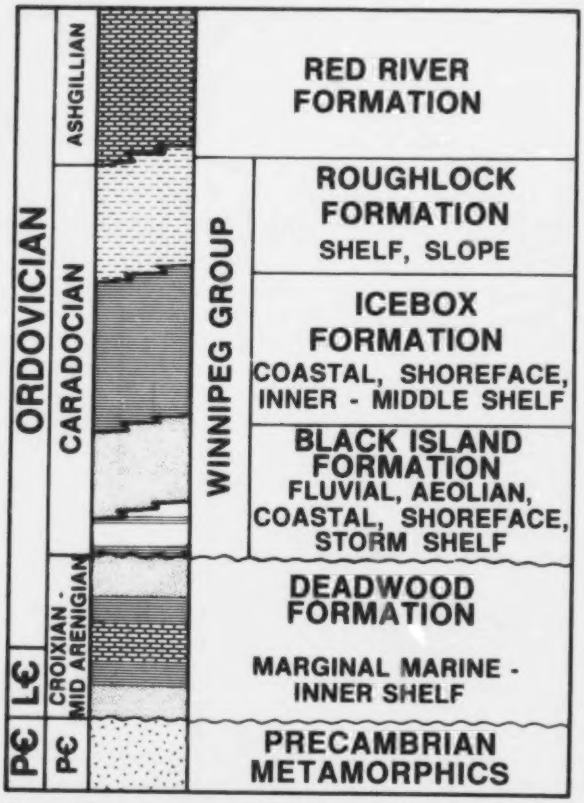

Figure 2. Precambrian - Upper Ordovician lithostratigraphic nomenclature, Williston Basin. Depositional environments for Deadwood Formation and Winnipeg Group are indicated.

more marine mudstone in the basin center. He also described a large (kilometers - 10's of kilometers in width and length) east-west trending sand body, about 100 km seaward of his upper Winnipeg paleo-shoreline in southern Manitoba. This sand body was first described by Andrichuk (1959) and is commonly referred to as the "Carman Sandstone." Vigrass (1971) suggested it to be an offshore bar formed by longshore currents with sediments derived from the east beyond the erosional edge of the Winnipeg. The environmental interpretations presented in this study will be centered on the "Carman Sandstone."

Other conclusions have not differed greatly from Vigrass's interpretation. Paterson (1971) in a largely stratigraphic study in Saskatchewan saw the Black Island Formation as shallow marine with coastal facies on the basin edges and the Icebox as more shaly and open marine. He described sand bodies in the Icebox as migrating toward the center of the basin or discontinuously toward the west. The Roughlock Member under the Red River Formation was generally eroded in Saskatchewan. Paterson (1971) also suggested that the thickness variation in the Icebox was related to basement fault adjustment. McCabe (1978) in a study of Winnipeg hydrocarbon potential in Manitoba interpreted both the lower and upper Winnipeg (Black Island and Icebox) as being mostly shallow marine, and the northward shift of marine conditions in the upper Winnipeg to reflect continued basin subsidence and northward transgression. He described the "Carman Sandstone" as an offshore bar with considerable depositional relief and restricted marine conditions shoreward of it. The "Carman Sandstone" was deposited in the lee of a Precambrian high to the east. At the end of Winnipeg deposition a marine transgression drowned northern and eastern sediment sources. This event heralded the transition to Red River carbonate deposition. Kessler (1978) in a largely diagenetic study suggested that the lower Winnipeg was largely regressive with fluvial and aeolian facies on the basin edges whilst the upper Winnipeg was transgressive.

Thompson (1984), Lefever *et al.* (1986), and Carlson and Thompson (1987) have described the depositional setting of the Winnipeg Group, especially the Icebox Formation, in North Dakota. According to Thompson (1984) the Icebox Formation was deposited in a shallow nearshore setting with northerly and easterly sand input. As based on several wells in eastern North Dakota, Grand Forks County, part of an east-west trending sand body similar to the "Carman Sandstone" was delineated and described as offshore bars. The presence of cross-bedded and burrowed sandstone bodies also indicated offshore current deposition. The Lefever *et al.* (1986) and Carlson and Thompson (1987) studies suggested a similar depositional history for the Icebox Formation; *i.e.*, a continued transgression and associated sea level rise during its deposition.

NOMENCLATURE

The Winnipeg Group formation names, Black Island, Icebox, and Roughlock (Fig. 2) have been retained in this study. These names were originally applied by Carlson (1960) in the North Dakota subsurface; the lower two members were retained by Paterson (1971) in Saskatchewan; and the overall tripartite system adopted by Thompson (1984), Lefever *et al.* (1986), and Carlson and Thompson (1987) in North Dakota. Vigrass (1971) and McCabe (1978) divided the Winnipeg into upper and lower members; Vigrass because he believed the existing members are diachronous; and McCabe made 50 percent thickness slices for the purpose of counting discrete sands. Though both these methods serve specific purposes, for this regional study, as based on well log correlations, depositional environment analysis, and stratigraphic sequence delineation, the tripartite formation scheme is more useful.

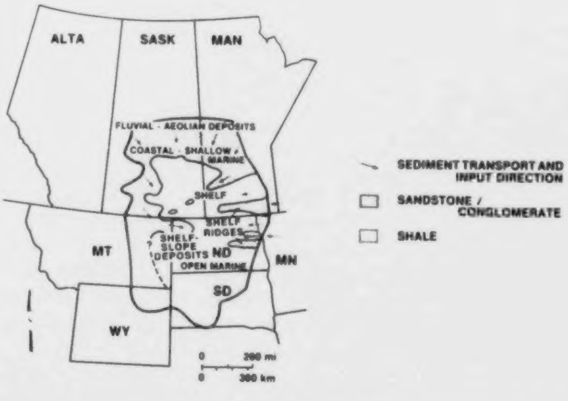

Figure 3. Depositional environment distribution and extent of sandstone facies in Icebox Formation (Winnipeg Group), Williston Basin.

METHODS OF STUDY

From the well logs through the Winnipeg in southern Manitoba, Saskatchewan, and NE North Dakota, three cross-sections (Figs. 8-10) and an isopach map (Fig. 4) have been constructed to show the geometry of and the relationship between the Carman sand in southern Manitoba and smaller equivalent elongate sand bodies in Manitoba, Saskatchewan, and North Dakota. Well sample and core descriptions (with representative photographs) from Manitoba and southern Saskatchewan are included. Comparisons with the geometry, sedimentary structure sequences, and paleogeographic position of other probable shelf ridges of storm and tidal origin in the Jurassic of the North Sea, and the Cretaceous of western Colorado are included.

Conodont zonation data of Oberg (1966) and Sweet (1982, 1984, pers. comm.) in the Winnipeg are analyzed along with Williston Basin subsidence data from Ahern and Mrkvicka (1984), Fowler and Nisbet (1985) and Crowley *et al.* (1985) in an attempt to understand the relationship of Icebox Formation shelf ridges to the Middle Ordovician transgression and subsidence. This relationship is discussed in terms of three defined stratigraphic sequences W1, W2, and W3.

"CARMAN SANDSTONE" RIDGE DEVELOPMENT

The Carman sand in southern Manitoba is an elongate east-west sand body which is 250 km long, 40-100 km wide, and up to 31 m thick (Fig. 4). It is encased in shale and mudstone of the Icebox Formation and is approximately 100 km seaward from an inferred Icebox shoreline as defined by Vigrass (1971) and McCabe (1978).

Though 26 wells penetrate the full thickness of the Carman sand none have been successfully cored due to the poorly consolidated nature of the sand. Evidence such as overall sand body shape and thickness, proximity to an interpreted coastline, and encasement in marine shale have been used for the interpretation of the Carman sand as an offshore bar built by currents that carried sand from sources now east of the present subcrop edge of the Icebox Formation. Of note is McCabe's (1978) observation of considerable depositional relief on the Carman sand body.

It is possible, however, using data on sand body thickness, well log shapes, and well sample descriptions to suggest a more detailed depositional format for the Carman sand. The distribution of sandstone thicknesses of >70.0 ft (21.3 m) in the center of the Carman sand body does not indicate a smooth ridge along the axis of the feature (Fig. 4), but rather, a complex pattern of offset zones of thickness migrated toward the south in an easterly direction. As shown in Figure 4, the first area on the west end of the Carman sand features thicknesses of 77.0 ft (23.5 m), 101.0 ft (30.8 m), and 83.0 ft (25.3 m) from west to east. The next area of thickness is slightly offset to the south and displays from west to east thicknesses of 85.0 ft (25.9 m), 82.0 ft (25.0 m), 86.0 ft

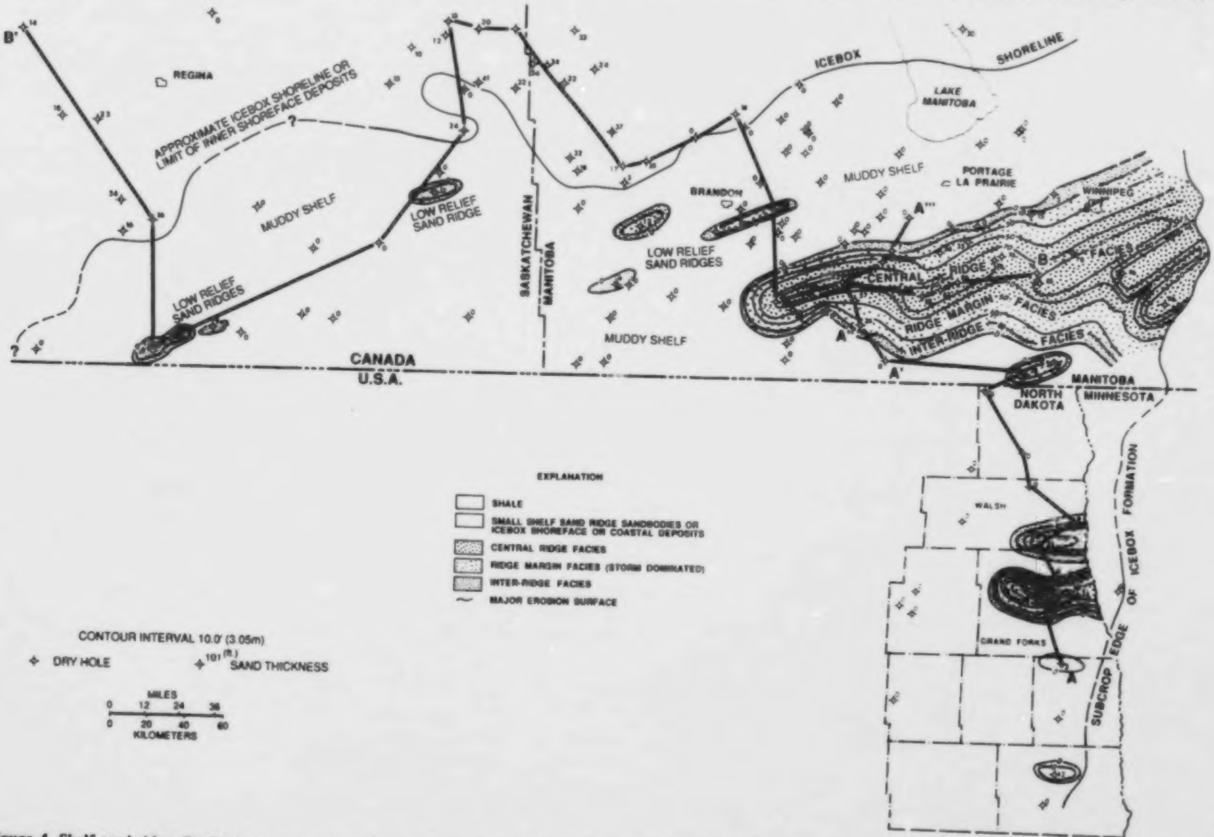

Figure 4. Shelf sand ridge distribution and sand isopach map, Carman sand and associated shelf ridge deposits, Icebox Formation (Winnipeg Group), northwest Minnesota, northeast North Dakota, southern Manitoba, southern Saskatchewan.

(26.2 m), and 76.0 ft (23.2 m), respectively (Fig. 4). Eastward toward the subcrop edge of the Carman sand the pattern repeats. The offsets indicate not one coherent feature, but at least 5 coalescing sand bodies which migrated to the north and the west.

Another interesting, though slightly more speculative, feature of the Carman sand is the apparently upward increase in grain size as observed in well samples and interpreted from spontaneous potential (SP) and gamma ray (GR) well log curves. In the 2-35-5-8W1 well where the Carman sand is 85.0 ft (25.9 m) thick (in the second thick east from the western end of the Carman sand) a generally coarsening upward from fine sandstone at the base through medium to medium-coarse sandstone through the middle and at the top of the sand body (Fig. 5) is seen in the samples. This distribution of Carman sand sizes is depicted on the SP. At the base of the log a lower SP increasing upward corresponds with the zone of fine sandstone observed in the well samples (Fig. 5). Above this, a more complex reading indicates a generally coarser sand (medium grained sandstone in the well samples). The higher SP readings at the top of the Carman sand correspond to the medium-coarse sandstone in the well samples (Fig. 5). A similar tripartite sandstone grain size distribution corresponding to changes in SP log character is observed in 16-22-7-10W1 well where the Carman sand is 53.0 ft (16.2 m) thick (Fig. 6). This well is on the north edge of the westernmost and most northerly thick in the Carman sand. On the southern edge of this area in the 16-13-5-12W1 well the Carman sand is 57.0 ft (17.4 m) thick and has a more complex pattern of coarsening upward sandstone grain size distribution, which is illustrated by the well sample description and the gamma ray log curve (Fig. 7). Of note are medium sandstone beds separated by coarse sandstone. Grain size distribution patterns of this sort seem to be peculiar to the edges of the thicks in the Carman sand.

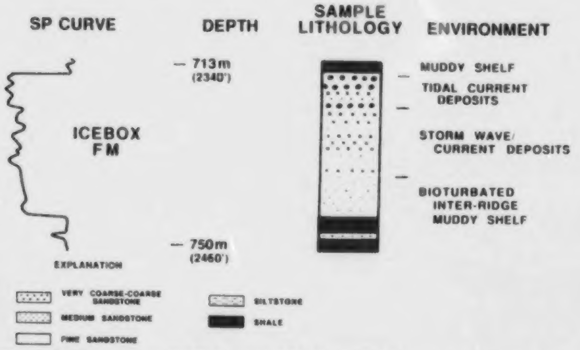

Figure 5. SP curve, sample lithology, and depositional environment interpretation, 2-35-5-8W1 well, Carman sand, southern Manitoba.

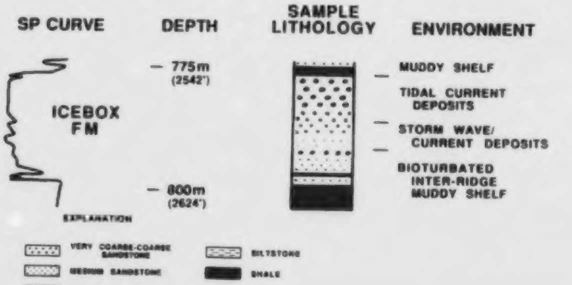

Figure 6. SP curve, sample lithology, and depositional environment interpretation, 16-22-7-10W1 well, Carman sand.

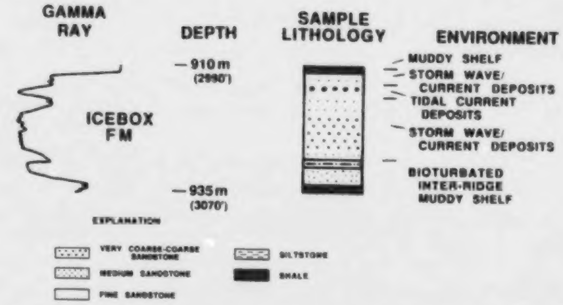

Figure 7. Gamma ray curve, sample lithology, and depositional environment interpretation, 16-13-5-12W1 well, Carman sand, southern Manitoba.

From the complex geometry of the Carman sand (its apparent division into at least 5 coalescing sand bodies, its upward division into 3 sequences of progressively coarsening sandstone) and its location about 100 km (60 miles) seaward of the Icebox coastline, a new interpretation can be made as to its depositional environment. Based on these factors the Carman sand is a series of coalescing and shoaling upward mid-outer shelf ridges which were formed in a three stage process. The initial two stages of deposition are thought to involve storm-driven geostrophic currents on the shelf which carried sand sub-parallel to the shoreline. The finer grained basal part of the ridge as based on Mesozoic outcrop and core examples, is thought to be thin-bedded and highly bioturbated. The slightly thicker bedded and coarser section above it is taken to be less bioturbated and to include remnants of storm derived bedforms such as hummocky cross-stratification (HCS). The coarser sand capping the ridge would be cross-bedded in decimetre - to metre lenses, not unlike those observed in tidal channels.

This may seem a rather detailed interpretation of a subsurface sand body which has not been successfully cored, but well documented depositional features of similar geometry and grain-size characteristics have been attributed to similar depositional settings. Several shelf sand bodies, like the Carman sand coalescing shelf ridges, have been described in outcrop from the Upper Cretaceous (Campanian) of northwestern Colorado by Boyles and Scott (1982) and Elliot (1987). The sands which coarsen upward, are 25-40 m thick (compared to 30 m maximum thickness in the Carman sand), 4-12 km wide (including subsurface equivalents), and perhaps 50 km or greater in length. They consist of a bioturbated, thin bedded, fine grained basal sandstone, grading upward to a fine-medium faintly bedded (planar and HCS) sandstone, and capped by medium coarse cross-bedded sandstone in metre-decimetre lenses.

A difference between the Upper Cretaceous sand ridges and the Carman sand is the interpreted distance from shore; about 25 km (15.5 miles) in the Duffy Mountain Sandstone, versus the 100 km (60 miles) in the Carman sand. A reason for this difference may be related to shelf subsidence rate and its effect on sediment supply and water depth farther than 25 km offshore, *i.e.* the ability of the sand ridge to shoal, aggrade, and migrate.

According to Boyles and Scott (1982) and Elliot (1987) storm-driven geostrophic currents were the major process in construction of the lower two sandy beds shelf ridges. Likewise, Boyles and Scott (*op. cit.*) attributed the top cross-bedded unit to storm-driven geostrophic current origin. However, Elliot suggested that strong offshore tidal currents were responsible and noted the similarity of the cross-bed sets (down to reactivation surfaces and mud drapes)

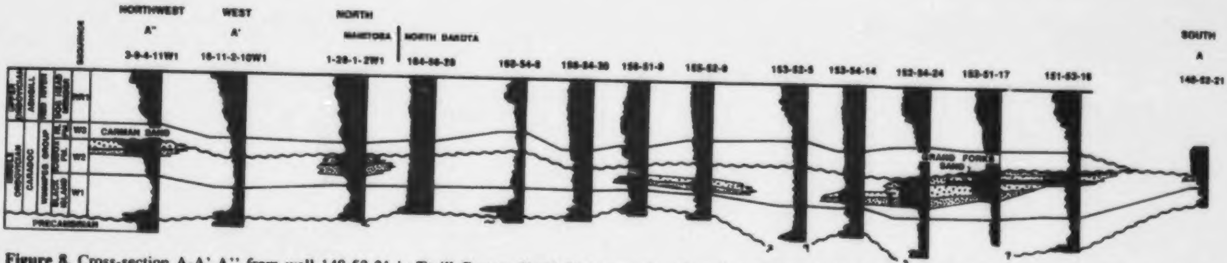

Figure 8. Cross-section A-A'-A'' from well 148-52-21 in Traill County, North Dakota north and northwest to wells 16-11-2-10W1 (A') and 3-9-4-11W1 (A'') in southern Manitoba. This cross-section correlates the Grand Forks sand in Grand Forks County, North Dakota, to the Carman sand in southern Manitoba.

to those observed in modern and ancient tidal deposits. In addition, he argued that the shelf ridges shoaled sufficiently to be influenced by such tidal currents. This suggestion fits well with McCabe's (1978) observation of considerable depositional relief on the Carman sand. This relief suggests aggradation of a sand ridge and is expressed in cross-sections A'-A''-A''' and B-B' through the Carman sand (Figs. 9-10).

Other subsurface examples of shelf ridges can be cited. Tillman and Martinsen (1984, 1987) described a shelf ridge complex in the Shannon Sandstone (Campanian) in Hartzog Draw Field of east-central Wyoming. Ridges up to 15 m (50.0 ft) thick, 6 km (4.0 miles) wide and 35 km (22 miles) long are observed. As in previous examples, they coarsen upward. From core information they do not differ greatly in sedimentary structure characteristics from the outcrop example of Elliot (1987). For descriptive purposes Tillman and Martinsen's (1984, 1987) terminology for naming the parts of a shelf ridge is used. The lower, finer, bioturbated part of the shelf ridge is referred to as the inter-ridge facies; the middle, medium-grained, storm-current dominated part is referred to as the ridge margin facies, and the coarser, tidal current part is referred to as the central ridge facies.

Thus the cross-section A'-A''-A''' which is south to north across the western end of the Carman sand shows the distribution of the inter-ridge, ridge-margin, and central ridge facies (Fig. 9). Of note is the strongly aggradational character (considerable depositional relief) of the Carman sand. The top of the central ridge facies is shown as being, in part, eroded. This feature will be discussed in the section on subsidence and sequence development. Cross-section B-B' lies east-west through the Carman sand. In addition to the considerable depositional relief observable, is the depth variation on the top of the Carman sand in wells 13-11-6-12W1 and 2-35-5-8W1 and 9-22-6-6W1 and 1-12-7-2W1 (Fig. 10). Instead of accounting for these differences by later faulting or differential subsidence, the author suggests that the edges of coalescing sand ridges occur between these wells. From the map of the Carman sand in Figure 4 this is shown to be a strong possibility. For this reason facies boundaries are included between individual sand ridges of the Carman sand on cross-section B-B' (Fig. 10).

OTHER SHELF RIDGE AND RELATED DEPOSITS

In the subsurface of Grand Forks and Walsh Counties, North Dakota, two smaller (10's of kilometers in length) elongate sand bodies, similar in geometry to the Carman sand, occur in the Icebox Formation (Figs. 3, 4, 8). These sand bodies are up to 58.0 ft (17.7 m) thick and from SP and GR log curves appear to generally coarsen upward in grain size in the main sand body of Grand Forks County (Fig. 4). Coarser sand at the top is referred to as central ridge facies, whereas finer sandstone toward the base of the feature are included in the ridge-margin and inter-ridge facies, respectively. The lower relief finer grained northern sand body which is mostly in Walsh County is interpreted as containing ridge margin and inter-ridge facies only (Figs. 4 and 8). As with the Carman sand, core data are not available.

Cross-section A-A'-A'' (Fig. 8) shows a complex relationship between individual sands of the main Grand Forks sand body. In the 152-51-17 and 151-53-16 wells the sand body splits into two sands, indicating two distinct phases of shelf ridge development with perhaps some lateral coalescing of small ridges. This complexity is further illustrated on the northern edge of the main Grand Forks sand body in the 153-54-14 well where a low relief finer grained (ridge margin and inter-ridge facies) appears to coalesce with the main sand body (Fig. 8).

As depicted in Figure 4, the two major Icebox sand bodies are about 10 km from the subcrop edge of the Icebox Formation in northwestern Minnesota. In fact, only a thin eroded shale of the Icebox remains in the well 152-49-11 in Polk County, Minnesota, about 8 km east of the main Grand Forks sand body. Thus it appears that only the western end of a much larger complex of coalescing sand ridges is preserved in this area.

The Grand Forks sand ridges can be correlated with the Carman sand about 140 km (85 miles) to the north (Fig. 8). Though the evidence is gone (beyond the subcrop erosional edge of the Icebox), they might both be part of a large northerly and westerly migrating system of upward shoaling shelf sand ridges, the lateral extent of which is dependent on vicissitudes of sediment supply, water depth, and the strength of geostrophic and offshore tidal currents.

Two thin sand accumulations [3.0 ft (0.9 m) and 12.0 ft (3.7 m), respectively] occur south of the main Grand Forks sand body. They may represent erosional remnants of other low relief sand ridges or small accumulations in the lee of a now eroded major sand ridge to the east.

MANITOBA

A series of thinner possibly elongate sand accumulations are also associated with the Carman sand (Fig. 4). The thickest of these is 33.0 ft (10.1 m) and is penetrated by the 1-28-1-2W1 well. This accumulation appears to coarsen, then fine upward and is interpreted as being composed of ridge margin and inter-ridge facies. It may represent an incipient ridge, which due to a lack of sediment supply in lee of a thick part of the Carman sand, never shoaled enough to encounter strong offshore tidal currents. The three other small sand accumulations to the northwest of the Carman sand in Manitoba may be of similar origin. It is possible that the elongate sand body just south of Brandon, Manitoba, with a maximum thickness of 16.0 ft (4.9 m) may represent the basal deposits of what could have been the next northwestward accreting ridge of the Carman sand (Fig. 4). A loss of sediment supply or a change in water depth due to basin subsidence or other basin hydrographic changes may have hindered its growth. As

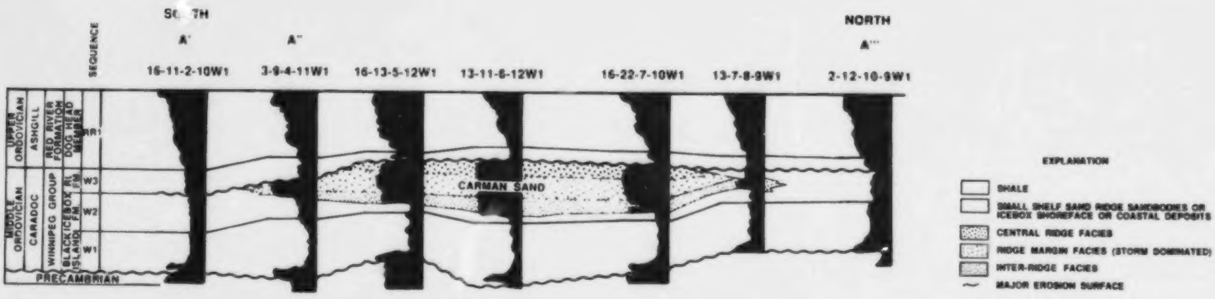

Figure 9. Cross-section A'-A''-A''' from well 16-11-2-10W1 north to well 2-12-10-9W1 in southern Manitoba. This cross-section cuts south-north across the Carman sand.

delineated in cross-section B-B' (11-9-10-16W1 well), this sand body is equivalent to the upper part of the most northerly sand ridge in the Carman sand, and thus was probably terminated by the erosional event at the top of the Carman sand (Fig. 10).

SASKATCHEWAN

To the west of the Carman sand other low relief, apparently elongate sand bodies are observed in the Icebox Formation in southern Saskatchewan (Fig. 4). These occur in the most shaly part of the formation in an area described as muddy shelf (Paterson, 1971; Vigrass, 1971).

The sediment source for these sand bodies may well be the same sort of storm driven geostrophic currents which were responsible for the Carman sand. Perhaps, during rare periods of heightened storm activity these currents reached further west than usual and formed elongate low relief sand accumulations. A lack of sediment supply (due to less geostrophic current activity or too great a distance from sediment supply) or an increase in a water depth (preventing shoaling or access to shallower tidal currents) prevented further sand ridge accretion and aggradation.

From cores, it has been determined that in the westernmost elongate sand body in southern Saskatchewan, 17.0 ft (5.2 m) of fine-medium grained planar and trough cross-bedded (rare planar bedding) sandstone (lying between two beds of highly burrowed fine sandstone) are encountered between 9877.0 ft (3010.4 m) and 9894.0 ft (3015.5 m) in the Icebox Formation in the Imperial Herald 1-31-1-20W2 well (Figs. 4, 10, and 11). Scour fill trough cross-beds with organic debris (mostly arthropod fragments) and shale rip-up clasts at their bases are observed throughout this part of the cored interval (Fig. 11). An example of scour fill trough cross-beds of this sort at 9881.3 ft (3011.6 m) is shown (Fig. 12a). Of note are oblique or opposing cross-beds at the base of the core slab. Such diversity of cross-bed orientation can occur in storm or tidal current deposits. Organic (arthropod fragments) and shale rip-up clasts are shown at the base of a scour fill at 9895.0 ft (3015.8 m) near the bottom of the cross-bedded part of the cored interval (Fig. 12b). The imbrication of small rip-up clasts suggests sediment transport by episodic strong tractive currents eroding the floor of the muddy shelf. The currents which deposited these bedforms could well have been storm driven and geostrophic. The absence of hummocky cross-stratification (HCS) which is often taken as an indicator of storm activity does not deny this interpretation. The formation of HCS is generally attributed to combined orbital (wave) and uni-directional current activity under storm conditions (Sherman & Greenwood, 1985). Perhaps wave effects were not deep enough to interact with storm current activity on this part of the shelf.

In Manitoba and Saskatchewan cross-section B-B' crosses areas shoreward of the Carman sand and other possible smaller shelf sand ridge deposits (Fig. 10). Here, highly burrowed fine sandstones with occasional cross-bedding are observed in core (*e.g.*, 1-4-20-32W1), and probably represent inner shelf/outer shoreface conditions which supply little sediment to the mid-outer shelf (Fig. 10).

Farther the north, in wells such as 9-32-20-27W2, well developed trough and planar cross-bedding which cut into thin zones of a burrowed fine sandstone and shale (Figs. 10, 13) occur. A scour surface near the base of the cored interval is shown at 6264.2 ft (1909.2 m) (Fig. 14a). These cross-bedded sandstones fine upward from coarse grained bases and may well represent gutter-cast or rip current channel fills in an outer shoreface setting. Such features have been observed from the subsurface Middle Jurassic of the Norwegian North Sea by Cockings *et al.* (1991) and in outcrop by Goldring (1971) in the Devonian of southwestern England and in the Upper Precambrian of Scotland by Kessler and Gollop (1988). Such channels feed sand to the inner shelf area, but not much farther.

A sharp erosional boundary is observed at the base of the Icebox Formation in the 9-32-20-27W2 well at 6265.8 ft (1909.7 m) (Fig. 14b). Here interbedded fine sandstone overlies a faintly cross-bedded sandstone of beach or upper shoreface origin at the top of the Black Island Formation. As will be discussed in the sections on subsidence and sequence development, this sharp boundary represents the shoreface equivalent of a basin edge flooding surface which marks the beginning of the transgression leading to deposition of the Carman and Grand Forks shelf sand ridges.

THE ORIGIN OF SHELF SAND RIDGES

There is general agreement that earliest subsidence in the Williston Basin began somewhere between the Late Cambrian and Late Ordovician. Based on isopach mapping and basin modeling techniques such as backstripping, Lefever *et al.* (1986) suggested that Williston Basin subsidence began during the Early Ordovician (Late Tremadoc). This is based on thickening trends observed on an isopach map of their "Member D" of the Deadwood Formation in the center (northwestern North Dakota) of what is now the Williston Basin. Isopach maps also showed that subsidence was well established and continuous by the Carodoc during which Winnipeg deposition occurred (Lefever *et al.*, 1986).This subsidence helps to explain the pattern of shelf sand deposition in the Icebox Formation.

The shelf areas in North Dakota, Manitoba, and Saskatchewan, where the Icebox sand ridges were deposited, represent the platform of the Caradoc Williston Basin where subsidence rates were probably low compared to those more basinward. Thus, major sediment input from west flowing geostrophic currents

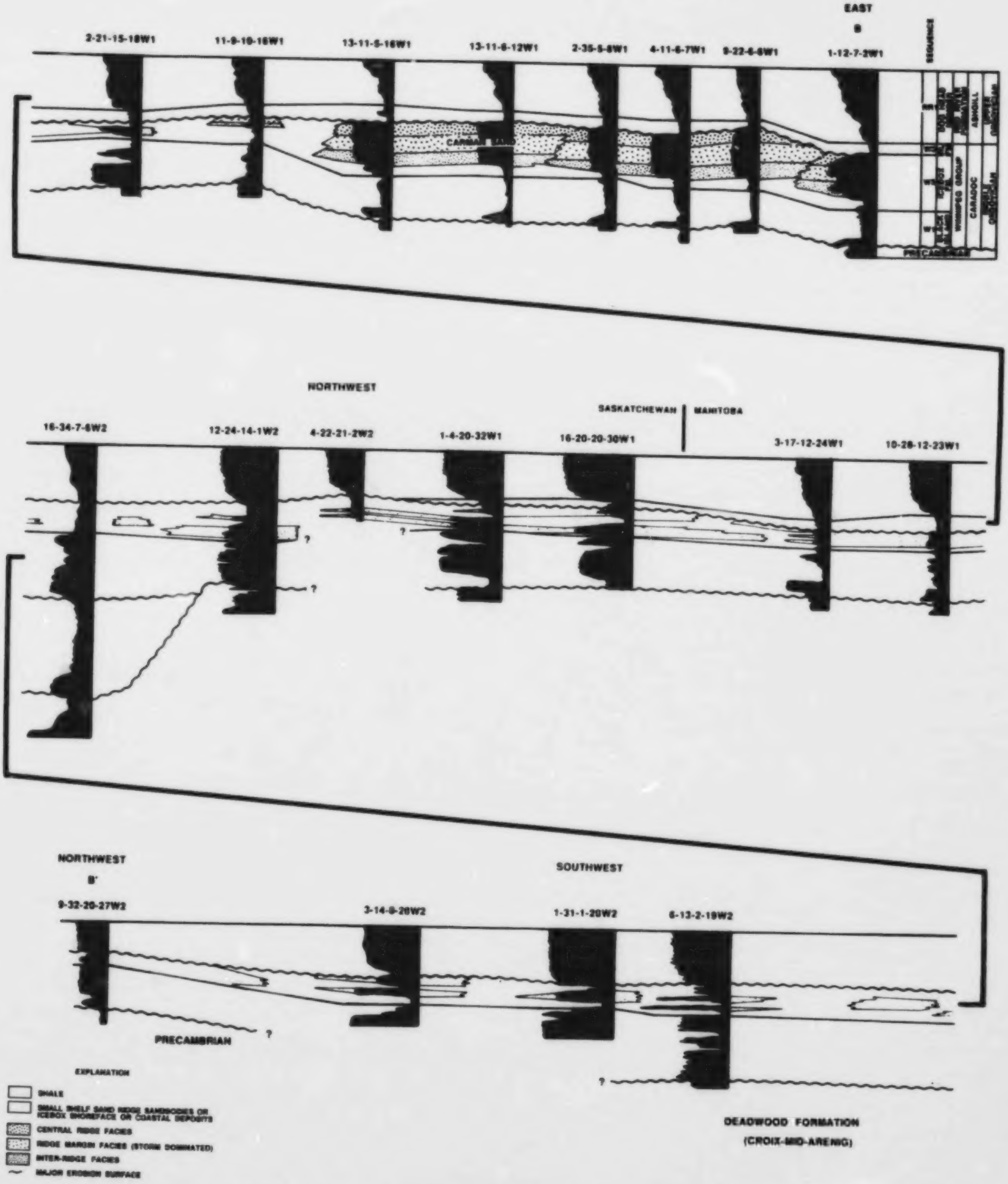

Figure 10. Cross-section B-B' from well 1-12-7-2W1 in the Carman sand in southern Manitoba west and northwest into southern and central Saskatchewan to well 9-32-20-27W2. Shows the relationship of the Carman sand and small sandbody development to the west.

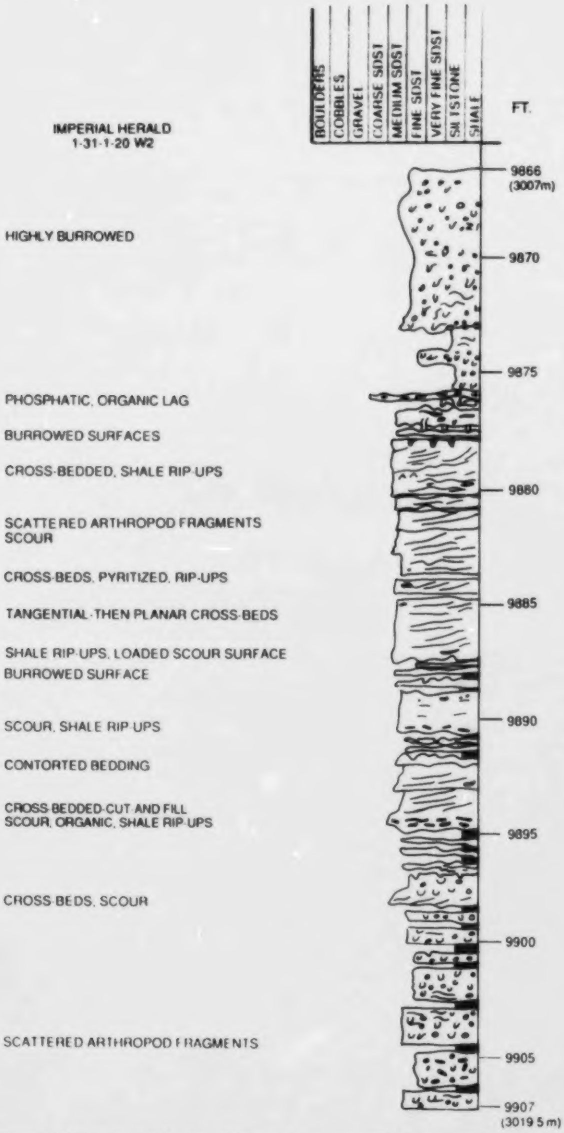

Figure 11. Core description log of Icebox Formation (9866.0 ft - 9907.0 ft, 3007.0 - 3019.5 m), Imperial Herald 1-31-1-20W2 well, southern Saskatchewan. Note cross-bedded interval between two burrowed zones.

coupled with a low subsidence rate allowed for the build up of sand ridges to a depth where offshore tidal currents might be effective. Increasing subsidence rates exceeding the ability of a ridge or ridge system to aggrade may have led to the abandonment of some, such as the Grand Forks sand body. In this instance, geostrophic currents built ridge systems such as the Carman sand farther to the north in Manitoba, where subsidence rates were lower.

Low relief sand ridges to the west of the Carman sand in the muddy shelf area of southern Saskatchewan may have been victims of more rapid basin subsidence in an area closer to the Williston Basin center. During periods of storm-driven geostrophic currents, these sediments were transported on to the muddy shelf to form low relief fine grained sand accumulations. However, as shown in the core from the Icebox Formation in the Imperial Herald 1-31-1-20W2 well, these sands are relatively thin (17.0 ft, 5.2 m) and capped by burrowed fine sandstone and shale (Fig. 11). These indicate an early end to ridge aggradation probably caused by subsidence rates in excess of sand influx.

Figure 12a. Low angle scour-fill cross-beds cutting into rippled and cross-bedded interval showing oblique or opposite transport direction. Canadian $1 coin for scale. 9881.3 ft (3011.7 m), Imperial Herald 1-31-1-20W2 well, southern Saskatchewan.

STRATIGRAPHIC SEQUENCES IN THE WINNIPEG GROUP

On a larger scale overall basin subsidence was likely a major factor in determining depositional patterns in the three formations. Fluvial and coastal deposits observed by such workers as Vigrass (1971), Kessler (1978), and Anderson (1984) in the Black Island

Figure 12b. Scour-fill sandstone will shale rip-up clasts and arthropod fragments at base. Note imbrication of small shale clasts in sandier upper part of core slab. Canadian $0.01 coin for scale. 9895.0 ft (3015.8 m). Imperial Herald 1-31-1-20W2, southern Saskatchewan

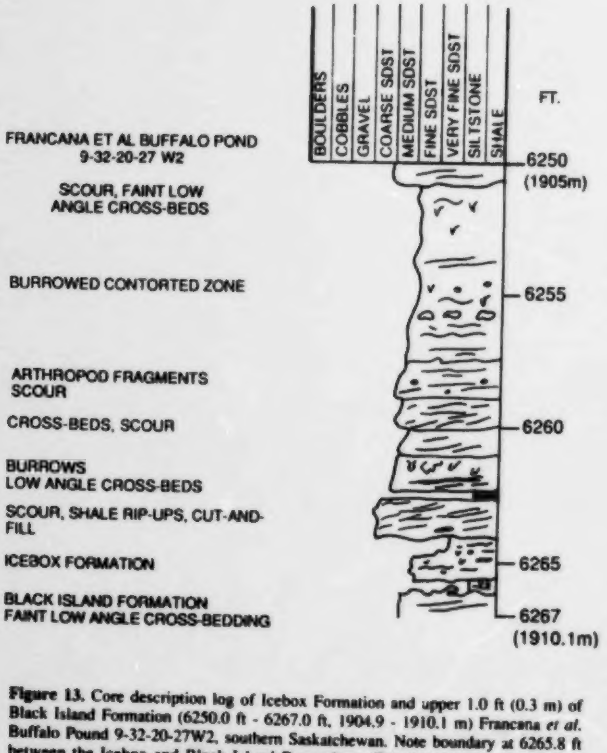

Figure 13. Core description log of Icebox Formation and upper 1.0 ft (0.3 m) of Black Island Formation (6250.0 ft - 6267.0 ft, 1904.9 - 1910.1 m) Francana *et al.* Buffalo Pound 9-32-20-27W2, southern Saskatchewan. Note boundary at 6265.8 ft between the Icebox and Black Island Formations. This boundary is the shallow marine flooding surface between sequences W1 and W2.

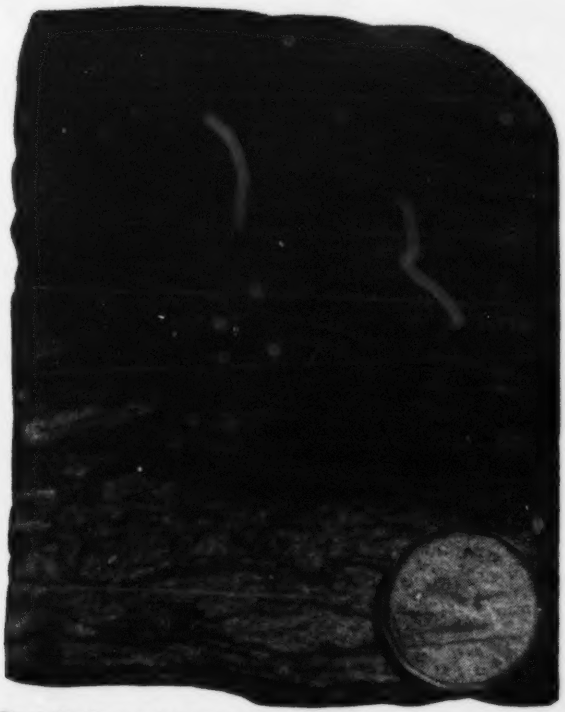

Figure 14a. Scour surface between cross-bedded medium-coarse sandstone and highly burrowed interbedded fine sandstone and shale near base of Icebox Formation, Canadian $1 coin for scale. 6264.2 ft (1909.2 m), Francana *et al.* Buffalo Pound 9-32-20-27W2, southern Saskatchewan.

Figure 14b. Slightly contorted, erosional contact between interbedded fine sandstone and shale of the Icebox Formation and the underlying faintly cross-bedded sandstone of the Black Island Formation. This contact marks the marine flooding surface which is the boundary between Winnipeg Formation sequences W1 (Black Island) and W2 (Icebox). Canadian $1 coin for scale. 6265.8 ft (1909.7 m). Francana *et al.* Buffalo Pound 9-32-20-27W2, southern Saskatchewan.

Formation became more marine upward with the increase in basin volume or accommodation space resulting from subsidence of even the platform surrounding the main basin. The culmination of this marine flooding marks the end of Black Island progradational deposition and the beginning of a transgressive/regressive cycle corresponding to the Icebox Formation.

For purposes of simplicity, generally contemporaneous stratigraphic units containing related depositional systems and separated from other such units by maximum marine flooding surfaces or in the instance of the Icebox and Roughlock Formations, a subsidence controlled unconformity, are herein referred to as sequences. These sequences are defined similarly to Galloway's (1989) "genetic stratigraphic sequences", with upper and lower boundaries at surfaces of maximum flooding (transgressions) and their basin equivalents. This format allows for the examination of basinward contemporaneous depositional systems (*i.e.*, fluvial, shoreface, sandy shelf, muddy shelf, *etc.*), the basic components of basin fill, and their boundaries with older and younger groups of similarly related depositional systems. Galloway (1989) applied this genetic stratigraphic classification to the Tertiary of the United States Gulf Coast and Cockings *et al.* (1991) described such sequences in the Middle Jurassic of the South Viking Graben in the Norwegian and British North Sea.

In the Winnipeg Group all three formations are referred to as sequences with the following designations: Roughlock Formation, W3; Icebox Formation, W2; and Black Island, W1. For the purposes of this study discussion of sequences will be limited mostly to W2 (Icebox Formation). The Black Island (W1) which is fluvial and coastal at its base and gradually becomes more marine upward and outward to the basin edges is a sequence with a maximum marine flooding surface separating it from W2, the Icebox Formation. The Roughlock Formation (W3), where it is observed in the study area, represents a subsidence controlled marine deepening (increase in basin accommodation space) which subdued the siliciclastic sedimentation of the Icebox Formation (W2). In areas of W2 sand deposition, such as the Grand Forks and the Carman sands and shoreface sands in Manitoba, this marine flooding surface manifests itself as an unconformity. In areas where it is not cut by erosion, sequence W3 (Roughlock) may be gradational with the overlying Dog Head Member of the Red River Formation and perhaps, the latter could be included in the same sequence.

The maximum marine flooding surface between W1 and W2 (the Black Island and Icebox) can be recognized in core and on well logs. In the 9-32-20-27W2 well (the most northwesterly well on cross-section B-B', Fig. 10) in southern Saskatchewan, a thin shaly zone above a coastal sandstone at top of W1 (Black Island) marks the shoreface equivalent of this maximum marine flooding surface. This is indicated by a rapid increase in gamma ray count at 6265.8 ft (1909.7 m) (Fig. 15). The gamma ray count quickly lessens with the shoreface sandstone just above this point. In core the upper surface of the Black Island sandstone is irregular and appears eroded with black shale containing scattered sand granules just above it (Fig. 14b). In a similar setting to the east in Manitoba in the 2-21-15-18W1 well, the flooding surface at the top of W1 occurs at a small limy deflection on the resistivity and SP logs within a shaly bed (Fig. 16). Above this point about 10 ft (3.0 m) of shale underlie a sand of probable shoreface or inner shelf origin. In the Carman sand, as illustrated by the 2-35-5-8W1 well, this same log deflection is interpreted as the flooding boundary between W2 and W1 (Fig. 17). Above this surface about 10.0 ft (3.0 m) of limy shale is overlain by 84.0 ft (25.6 m) of the Carman sand. A similar relationship is observed farther south in the Grand Forks sand where about 30.0 ft (9.0 m) of limy shale overlie the marine flooding surface (separating W2 and W1). This surface is picked by correlation of the same log deflection as in the Carman sand and northward (Fig. 18). Above the limy shale lies 58.0 ft (17.7 m) of Grand Forks sand.

The erosional surface at the top of the Carman and Grand Forks sands (Figs. 17 and 18) is thought to represent marine flooding when subsidence slowed and extra-basinal effects such as a general Caradoc sea level rise affected the edges of the Williston Basin (Barnes, 1984).

As stated, a sequence (between flooding surfaces) is a generally contemporaneous stratigraphic unit of related depositional systems. W2 (Icebox Formation) certainly meets these criteria with the spread of generally contemporaneous and related shoreface and muddy and sandy inner and mid shelf deposits. However, basin margin (fluvial deltaic, coastal, *etc.*) deposits are not preserved. They probably existed beyond what is now the subcrop edge of the Icebox Formation and were eroded due to the lack or small amount of subsidence on the Williston Basin margins.

The question of how internally contemporaneous are the elements of the W1, W2, and W3 sequences cannot be adequately addressed at this time. In their definition Cockings *et al.* (*op. cit.*) constrained their sequences with detailed palyno-stratigraphy. Boundaries between sequences though on occasion slightly diachronous were not inconsistent with biozonal boundaries. Conodont studies of the Winnipeg Group are scattered in areal extent and do not present a coherent biostratigraphic framework for application over the basin. However, there are indications that sequences W1, W2, and W3 are not strongly diachronous. From a conodont zonation done from six stratigraphic test wells (H-1 to H-6) by W. C. Sweet for S. C. Thompson (1984), the boundary between North American conodont faunal zones 8 and 9 (Fig. 19) seems fairly consistent with the W1/W2 flooding surface, even in the more distant 152-51-17 well. The Grand Forks sand seems to lie consistently within faunal zone 9, though the top is eroded. Much of zone 9 and all of zone 10 are missing between the eroded top of the Grand Forks sand (W2) and the more marine Roughlock Fm (W3). This represents a gap of 4.5 million years, certainly a significant hiatus. It would be interesting to see how much of zones 9 and 10 are missing above the Carman sand body and in other places within the study area.

Oberg's (1966) study of conodonts from outcrops of the Icebox Formation in Lake Winnipeg reveals the presence of *'Oistodus' venustus* and *Panderodus panderi* which first appear at or above the boundary between zones 8 and 9 in the wells studied in Grand Forks County, North Dakota. These data suggest the possibility of a reasonable biostratigraphic correlation of the Black Island Formation (W2) between the Grand Forks area and southern Manitoba. Beyond this, the contemporaneity of W2 or the other sequences cannot be tested at this time.

CONCLUSIONS

The principal conclusions of this study are:

1. The Carman sand of southern Manitoba, the Grand Forks sand of North Dakota, and an unnamed sand body just north of the Grand Forks sand in North Dakota are mid-shelf sand ridges (or sand ridge systems) initially formed by westward moving, storm-driven, shelf geostrophic currents. These ridges or ridge systems are up to 30 m thick.
2. As these ridges aggraded to shallower depths, a top coarse sandstone unit (with probable decimetre-metre thick sets of trough cross-bedding) was deposited by offshore tidal currents.

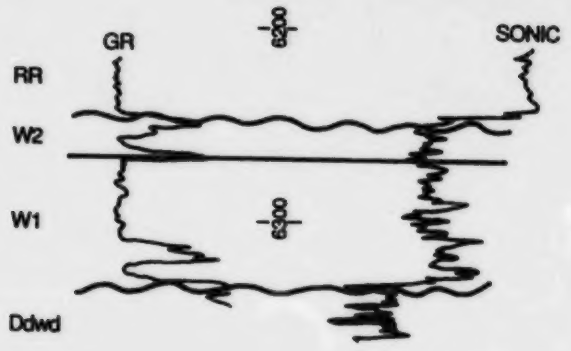

Figure 15. Stratigraphic relationship of sequences W1 (Black Island) and W2 (Icebox) in the Winnipeg Group, Francana *et al.* Buffalo Pound 9-32-20-27W2, west-central Saskatchewan. W3 is not observed in this part of Saskatchewan. W2 has thinned greatly from southern Saskatchewan as it approaches its subcrop erosional edge.

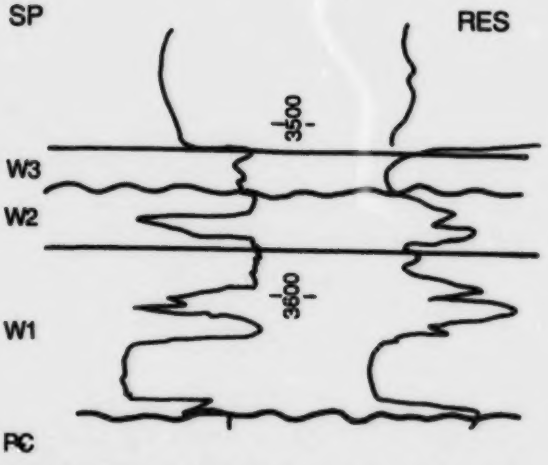

Figure 16. Stratigraphic relationship of sequences W1 (Black Island), W2 (Icebox), and W3 (Roughlock) in the Winnipeg Group, Gridoil Minnedosa 2-21-15-18W1, southern Manitoba. Boundary between sequences W1 and W2 is marine flooding surface; between W2 and W3 boundary is erosional. Sequence W2 has thinned greatly to the north of the Carman sand.

3. Fully developed shelf ridges in the Carman and Grand Forks sands are tripartite in nature consisting of a lower fine sandstone, a medial, fine to medium sandstone and a coarse sandstone unit at the top. Based on comparisons to Cretaceous outcrop and subsurface examples, these three units probably represent respectively:
   a) A highly burrowed fine grained sandstone.
   b) A fine-medium grained burrowed sandstone with remnants of storm induced bedding, such as hummocky cross-stratification (HCS).
   c) Medium-coarse grained sandstone with decimetre-metre size trough cross-bedding of possible tidal current origin.

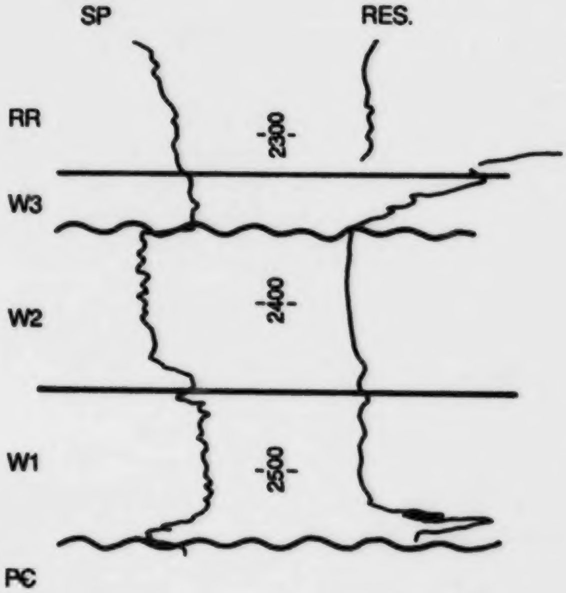

Figure 17. Stratigraphic relationship of sequences W1 (Black Island), W2 (Icebox), and W3 (Roughlock) in the Winnipeg Group, Sweetgrass Altamont 2-35-5-8W1, southern Manitoba. This well is in the Carman sand. Boundary between sequences W1 and W2 is marine flooding surface; between W2 and W3 boundary is erosional.

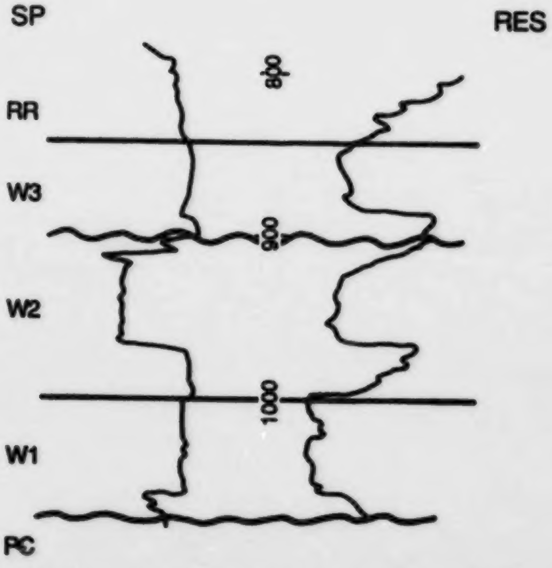

Figure 18. Stratigraphic relationship of sequences W1 (Black Island), W2 (Icebox), and W3 (Roughlock) in the Winnipeg Group, North Plains F.F. Danner #1 152-54-24SWSE, Grand Forks County, North Dakota. This well is in the Grand Forks sand. Boundary between sequences W1 and W2 is marine flooding surface; between sequences W2 and W3 boundary is erosional.

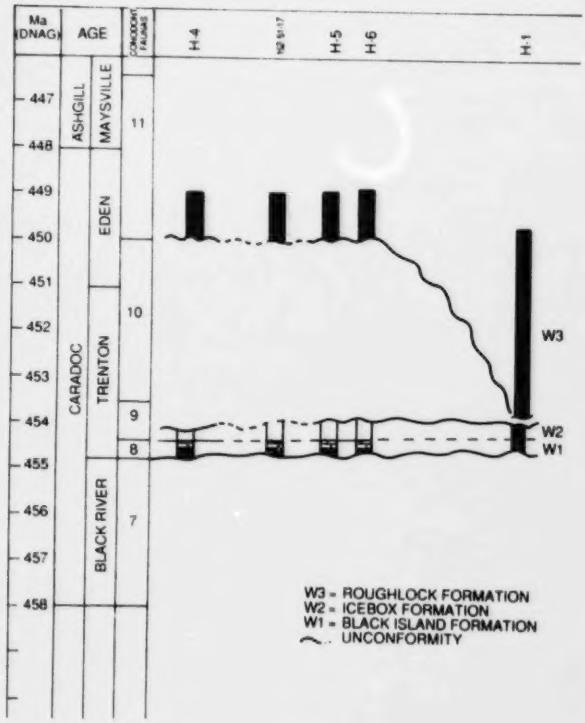

Figure 19: Conodont faunal group distributions and sequences boundaries in the Winnipeg Group in wells H-1, H-4, H-5, H-6 and 152-51-17, Grand Forks County, North Dakota. Note the missing conodont zones between sequences W2 and W3 in all wells except H-1 which is far to the south in Barnes County.

4. Small low relief elongate sand bodies of the Black Island Formation in southern Manitoba and Saskatchewan probably formed in the lee of the Carman sand shelf ridges during periods of exceptional storm-driven geostrophic current activity. These sand accumulations did not aggrade appreciably due to their location near the rapidly subsiding center of the early Williston Basin and/or a lack of constant sediment supply.

5. Contouring of the Carman sand indicates north and westward migration of individual coalescing sand ridges. This may be the result of slow subsidence of the shelf on the margins of the Williston Basin. Based on this reasoning, the Grand Forks sand may represent a slightly older and more basinward shelf ridge system which was abandoned due to increased basinal subsidence rates relative to sand input.

6. Based on the recognition of the marine equivalents of a basin margin flooding surface between the Black Island and Icebox Formations, the Winnipeg Group can be divided into three apparently contemporaneous stratigraphic sequences (W1, W2, W3) each containing related depositional systems. These sequences probably represent transgressive/regressive cycles controlled by changes in rates of basin margin subsidence.

7. Insufficient conodont biostratigraphic control of the Winnipeg Group exists for the determination of contemporaneity in the stratigraphic sequences. However, limited data show that the W1/W2 sequence boundary or marine flooding surface is contemporaneous in Grand Forks County, North Dakota.

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DIAGENESIS OF THE WINNIPEG FORMATION IN MANITOBA: A REGIONAL PETROGRAPHIC STUDY

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ABSTRACT

Despite the widespread extent of the Winnipeg Formation in the Williston Basin and its importance as a source of silica sand in outcrop and a potential oil and gas reservoir in the subsurface, surprisingly little is known of the petrography and diagenesis of this Ordovician siliciclastic unit. The Black Island Member, the lower coarse-grained clastic unit of the formation, was sampled from subsurface core and outcrop in Manitoba, and examined using thin sections, X-ray diffraction, and scanning electron microscopy. A surprisingly diverse suite of authigenic minerals exists, including: quartz, K-feldspar, Na-Ca plagioclase, Ca-plagioclase, anhydrite, calcite, dolomite, ankerite, halite, illite, kaolinite, and pyrite. Although there is considerable regional variation in southern Manitoba, the generalized paragenetic sequence recognized is: 1) Quartz formation as encrusting and terminated (euhedral) overgrowths; 2) Formation of anhydrite cement and neoformed crystals; 3) Precipitation of euhedral, 5 to 50 µm-sized feldspars; 4) Formation of anhydrite cement and neoformed crystals; 5) Dolomite precipitation; 6) Dedolomitization and formation of calcite crystals and cement; 7) Feldspar dissolution and clay mineral authigenesis; and 8) Formation of pyrite. These results show that the history of pore water solutions leading to the development of this diagenetic mineral suite is complex.

INTRODUCTION

The Ordovician Winnipeg Formation extends over central and southern Saskatchewan, the eastern region of Montana, all of North Dakota, and central and southwestern Manitoba. It is generally taken to be equivalent to the widespread Middle Ordovician cratonic sheet sands of the midcontinent region (*e.g.*, Saint Peter Sandstone). In Manitoba, the Winnipeg Formation crops out on several islands in Lake Winnipeg and is exposed on the southeastern shore of the lake (Fig. 1).

The term "Winnipeg Sandstone" was first used by Dowling (1895) in descriptions of the siliciclastic section underlying the Red River Formation carbonates and overlying the Precambrian metamorphic rocks along the shores of Lake Winnipeg. Since then, the nomenclature and stratigraphic subdivision of the Winnipeg "unit" has had a controversial history (*c.f.*, Fuller, 1961; Vigrass, 1971; Paterson, 1971; Sweet, 1982; Carlson and Thompson, 1987). Although variously named and complexly subdivided both formally and informally, in Manitoba the Winnipeg Formation is generally divided into two operational units: a lower coarse- to fine-grained sandstone "member" and an upper siltstone to shale unit. The formation overall thickens from less than 10 m in the outcrop belt at Lake Winnipeg to more than 60 m in the southwest corner of Manitoba (Andrichuk, 1959), and up to a maximum of about 135 m in the Williston Basin (LeFever *et al.*, 1987; Anderson, 1982). In most of Manitoba, the formation directly overlies the peneplained surface of the Precambrian rocks; the formation has a gradational upper contact with the limestone and dolostone of the Ordovician Red River Formation.

Despite differences of opinion regarding the detailed, formal stratigraphic nomenclature of the Winnipeg Formation, most investigators agree that the unit in Manitoba represents deposition in a transgressive, marine shoreline to offshore marine setting.

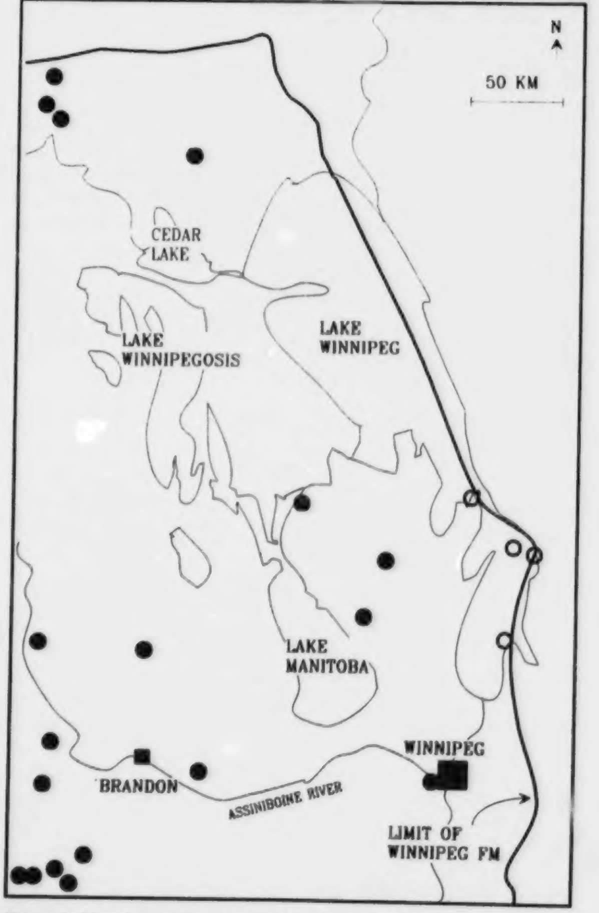

Figure 1: Map showing the location of samples used in this study. Open circles indicate outcrop samples; closed circles indicate subsurface core samples.

McCabe (1978) concluded that his Lower Winnipeg unit was deposited under high-energy, near-shore, shallow marine conditions. Northward trangression of the sea resulted in deposition of the finer grained Upper Winnipeg unit in more offshore marine environments under lower energy and deeper water conditions.

The lower, relatively coarse clastic unit, termed the Black Island Member (Genik, 1952), reaches a maxiumum thickness of 30 m in the southwestern corner of Manitoba. Outcrops of the unit have been quarried at Black Island as a source of silica sand since the mid-1950s. Despite the unit's wide geographic extent, and its importance as a potential oil and gas reservoir in the subsurface and as a source of silica sand in outcrop, relatively little is known of its petrography and diagenesis. The objective of this paper is to summarize the initial results of an on-going petrographic study of the Black Island Member in Manitoba.

METHODS OF STUDY

All available subsurface cores of the lower unit of the Black Island Member in Manitoba were examined (Fig. 1 and Table 1). Although the unit is not homogeneous in the cores, no attempt was made in this reconnaissance study to evaluate stratigraphic variations within the Member.

Epoxy-impregnated petrographic thin sections were made and examined using cathodoluminescence microscopy and transmitted-light microscopy. Thin sections and chips of approximately $1\ cm^3$ size were examined using scanning electron microscopy coupled with energy-dispersive X-ray and backscattered electron analysis. Mineralogy of the bulk, size segregated, and heavy mineral fractions were semi-quantitatively determined by X-ray diffraction. Details of these analytical techniques can be found in Shum (1990).

Table 1: Location of Core and Outcrop Samples.

Well/Outcrop Name	Location
Home *et al* S. Pierson	16-08-02-29WPM
Home Scurry S. Pierson	06-19-02-29WPM
Newscope Coulter #1	06-19-02-27WPM
Coulter	16-16-01-27WPM
MMR Waskada #6	11-29-01-25WPM
Chevron Reston #2	05-33-07-27WPM
Cal Stan Daly #23	15-18-10-27WPM
Arco Shlilo Prov #1	10-24-09-16WPM
Winnipeg Central Gas #4	13-10-02EPM
Imperial Madeline #1	16-18-18-29WPM
Dome Stratchair #5	08-34-16-21WPM
Bral Eriksdale #1	16-20-21-05WPM
Pascar #2	NE-11-25-08WPM
Bralorne Hilbre #1	08-30-29-08WPM
M-10-88	10-58-18WPM
M-17-86	09-60-28WPM
M-18-86	21-61-28WPM
M-23-85	32-62-27WPM
Victoria Beach, Lake Winnipeg	29-20-07EPM
Black Island, Lake Winnipeg	14-27-08EPM
Grindstone Pt, Lake Winnipeg	30-28-06EPM
Manigotagan	06-25-09EPM

RESULTS

GENERAL PETROGRAPHY OF THE BLACK ISLAND MEMBER

The Black Island Member of the Winnipeg Formation in Manitoba is a moderate to well sorted, fine to coarse grained, friable to poorly cemented sandstone. At the type locality on Black Island, the unit is a white to yellow, remarkably clean, loose to friable quartz arenite (sandstone terminology according to Dott, 1964, as modified by Pettijohn, 1975). In the outcrop, quartz commonly forms greater than 98 percent of the framework grains. Genik (1952) reports chemical analyses ranging from 95.2 to 99.4 percent silica. Bulk samples from the silica sand quarry on the Island are generally in excess of 98 percent SiO_2.

Although the quartz-rich nature of the basal sandstone unit in the outcrop area at Black Island is well known and documented, it is clearly incorrect to characterize all of the Black Island Member in Manitoba as a quartz arenite. In most of the samples examined from both subsurface and other outcrops, the Black Island Member is most commonly a quartz wacke with gradations from quartz arenite to subarkose, arkose, and arkosic wacke. Table 2 and Figure 2 show the ranges and average mineralogical composition of the unit.

The framework particles are usually dominated by well rounded and equant, monocrystalline and polycrystalline, detrital quartz (average 75 percent of the total rock). There are minor amounts of both detrital and authigenic feldspar minerals, igneous rock fragments, and heavy minerals. The most abundant feldspars present are orthoclase and microcline with smaller proportions of plagioclase varieties. Many of the K-feldspar grains are twinned. The heavy mineral content is dominated by pyrite, tourmaline, diopside, and zircon. Cements make up an average of about 18 percent of the rock, most of which are either carbonate minerals or anhydrite. Several investigators have identified and discussed a widespread pyrite-cemented zone and pyrite concretion/oolite bed that occurs near the top of the Black Island Member in both outcrop (Baillie, 1952; Genik, 1952) and subsurface cores (Macauley, 1952; Paterson, 1971; Binda and Simpson, 1989), but this pyritic zone was not sampled in the present study.

Table 2: Mineralogical Composition of the Black Island Member Sandstones in Manitoba.

	Mean percent	Range
Whole-rock Mineralogy (based on XRD analysis of 42 samples)		
Quartz:	74.9	42.6-99.0
K-feldspar:	2.8	0-31.3
Plagioclase:	0.7	0-4.1
Anhydrite:	5.0	0-39.6
Dolomite:	4.9	0-12.1
Calcite:	1.1	0-45.1
Illite:	7.9	0-31.9
Kaolinite:	0.2	0-4.6
Pyrite:	1.6	0-18.5
Halite:	<0.1	0-2.2
Framework Components (based on thin section analysis of 42 samples)		
Quartz:	90.9	64.6-99.5
Feldspars:	5.0	0-29.9
Rock Fragments:	1.8	0-5.9
Opaques:	2.5	0-12.6

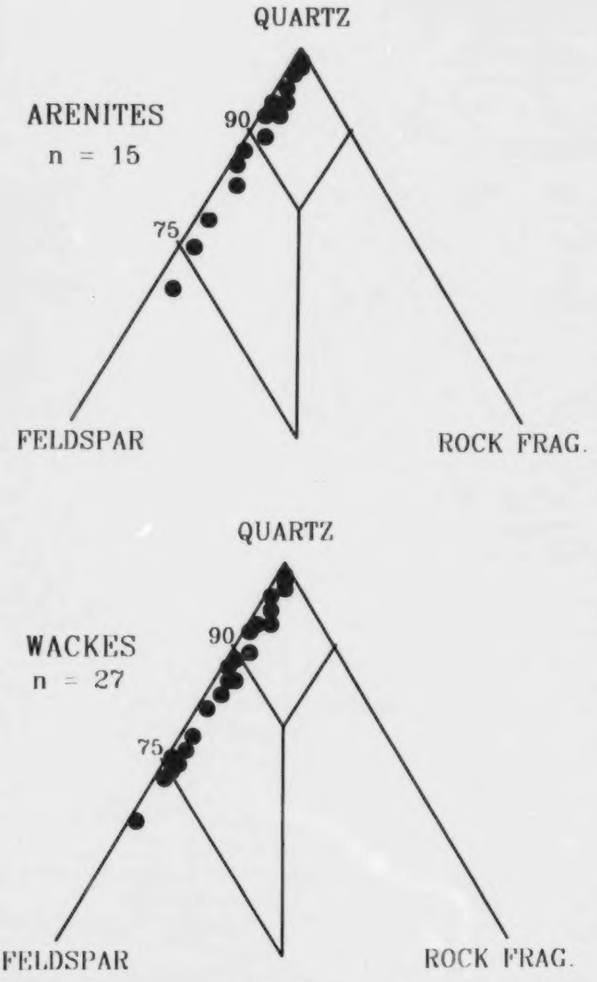

Figure 2: Triangular composition diagrams for the Black Island Member sandstones examined.

The fabric of the Black Island sandstone is typical of arenites. Most of the rock is supported structurally by the equant framework grains. Grain contacts range from tangential to concavo-convex, although extensive quartz overgrowths in some samples tend to mask them. Floating and point-contact grain fabrics are common in the southwestern part of the study area where poikilotopic anyhydrite cements have preserved the original loose packing of the grains (Fig. 3). A large proportion of the framework grains is rounded to well rounded. However, distinction between true detrital grains and grains the shapes of which have been modified by secondary mineral overgrowths is difficult in normal thin-section examination.

DIAGENESIS OF THE BLACK ISLAND MEMBER

Cementation of the Black Island sandstone is variable. In outcrop the unit ranges from loose, essentially uncemented sand to hard, well-indurated sandstone, and cements are mainly quartz and calcite. The quartz cement occurs as incipient and encrusting overgrowths on most of the detrital grains. The calcite cements are most commonly poikilotopic. Although several investigators have also noted the presence of iron oxide and iron sulphide-cemented laminae and beds in exposures of the Lake Winnipeg region (*e.g.*, Macauley, 1952; Genik, 1952), only very minor Fe-cementation was noted in the outcrop samples collected in this study.

Poor recovery in several of the cores indicates a lack of cementation in regions of the subsurface. Where the Black Island is somewhat more lithified, cements normally comprise about 10 to 30 percent of the total rock volume, and consist of (in approximate order of abundance): anhydrite, carbonate minerals, clay minerals, and quartz.

Anhydrite cement is most abundant in the southwestern part of the study area where it can compose as much as 40 percent of the rock. It occurs as large, lath-shaped, pore-filling crystals and as small, micron-sized euhedral neoformed crystal aggregates (Fig. 4). Although small amounts of authigenic halite are sometimes associated with calcium sulphate, when present, anhydrite is normally the dominant cement in the sandstone.

Carbonate minerals also commonly occur as pore-filling cements and as euhedral neoformed crystals in the subsurface sediments. The carbonates range from less than 5 percent to more than 30 percent of the rock. Dolomite and calcite cements are dominant, but ankerite is also present. Under cathodoluminescence the carbonates show rather complex growth relations. In the southern-most part of the study area, reddish-orange luminescent, euhedral dolomite rhombs often have a yellow rim of calcite, suggesting dedolomitization, *i.e.*, replacement of dolomite by calcite. In addition, Macauley (1952) noted evidence of dolomitization of pre-existing calcite cement in some outcrop samples.

Throughout the Black Island Member in the subsurface feldspars occur as euhedral to subhedral crystals ranging from 5 µm to 50 µm in size (Fig. 5). In some samples the feldspar content exceeds 30 percent of the total rock. Both sodic and potassic feldspars are present but monoclinic orthoclase is the most common. Although the feldspars are not cementing agents *per se*, they are clearly postdepositional neoformed minerals. Many of the subhedral feldspar crystals show some degree of alteration and transformation to clay minerals, usually flaky and fibrous illite. Small amounts of kaolinite have also been identified on the surfaces and in alteration clusters on the feldspar crystals. Authigenic illite can also be found coating the secondary quartz overgrowths on the detrital grains (Fig. 6).

In contrast to the observations by Macauley (1952, p. 37) and others (*e.g.*, Wallace and McCartney, 1928; Andrichuk, 1959) that secondary quartz overgrowths are rare in the Winnipeg sandstones, we note that quartz overgrowths are ubiquitous in subsurface samples. However, in many cases it is, indeed, difficult to recognize the overgrowths without the aid of cathodoluminesence and scanning electron microscopy. Although abundant in all samples, encrusting and incipient overgrowths are the most commonly observed forms (Fig. 7). Terminated overgrowths showing well-developed rhombohedral crystal faces are present but less common (Fig. 8). In most samples examined the quartz overgrowths are not interlocking.

Pyrite occurs in small amounts in many of the samples studied. Like the authigenic feldspar crystals, it is a neoformed mineral occurring in a wide variety of morphologies, including individual crystals and clusters of pyritohedra, octahedra, and framboids.

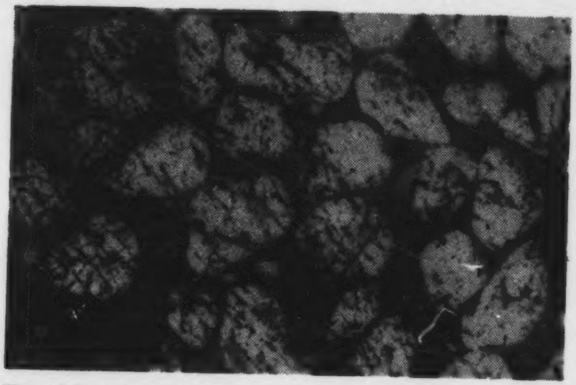

Figure 3: Photomicrograph of floating and point contact framework grain fabric in Black Island sandstone (field of view is 2.8 mm). Sample from 191.2 m depth in Pascar #2 well.

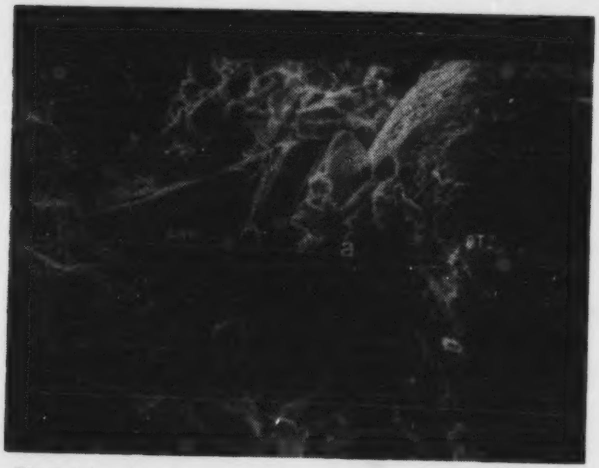

Figure 4: Scanning electron photomicrograph showing both coarsely crystalline, pore-filling anhydrite cement (ANH) and smaller, neoformed anhydrite crystal clusters (a); detrital quartz grain is labeled QTZ. Sample from 1683.0 m depth in Cal Stan Daly #23 well.

Figure 5: Scanning electron photomicrograph showing abundant euhedral authigenic K-feldspar (K). The feldspars postdate both the encrusting quartz overgrowth (Q) and the anhydrite cement (A). Sample from 1350.0 m depth in Imperial Madeline #1 well.

Figure 6: Scanning electron photomicrographs showing authigenic flaky illite (I) forming on the surface of a detrital quartz grain (q) and on the euhedral terminated quartz overgrowth (Q). Image on the right is a 4x enlargement of the highlighted area on the left. Sample from 255.0 m depth in Bralorne Hilbre #1 well.

Figure 7: Scanning electron photomicrograph showing encrusting quartz overgrowth (Q). Sample from 1683.0 m depth in Cal Stan Daly #23 well.

Figure 8: Scanning electron photomicrograph showing terminated quartz overgrowth (Q). Sample from 258.0 m depth in Bralorne Hilbre #1 well.

SPATIAL VARIATION AND PARAGENESIS OF DIAGENETIC MINERALS

The Black Island Member of the Winnipeg Formation shows considerable spatial variation in terms of the authigenic mineral assemblage and, hence, the interpreted diagenetic evolution of the rock. Although these results must still be considered preliminary, at least four separate regions can be recognized in southern Manitoba on the basis of the suite of diagenetic minerals present. These regions also exhibit somewhat different paragenetic sequences of authigenic minerals and cements, as interpreted from the microtextural and morphological relationships revealed by thin section and SEM examination. The four regions are shown in Figure 9 and summarized in Table 3.

Table 3: Summary of Sandstone composition of the Black Island Member in the Four Diagenetic Regions of Southern Manitoba.

	A	B	C	D
Framework components				
Quartz	69	77	82	70
Feldspars	8	8	1	1
Rock Frag.	1	0	0	0
Opaques	2	1	2	1
Cements and matrix				
Anhydrite	15	1	0	0
Carbonates	5	11	7	4
Clays	1	1	8	25
Quartz	1	2	2	1

Region A

The Black Island Member in the southwestern part of the province (Region A) contains the most complex diagenetic mineral suite, consisting of K-feldspar, Na-Ca and Ca-plagioclase, illite, anhydrite, carbonate minerals, quartz, halite, and pyrite. This region has the lowest proportion of detrital quartz grains and most of the sandstones are arkose and subarkose. Most of the feldspars show some degree of alteration. Cements are mainly calcium sulphate with only minor carbonate.

The sequence of diagenesis in this region is as follows: 1) precipitation of Na-Ca and Ca-plagioclases; 2) minor quartz overgrowths encrusting authigenic feldspars and detrital quartz grains; 3) precipitation of pore-filling anhydrite cement; 4) formation of K-feldspar; 5) pitting and etching of the potassic feldspars and the presence of illite associated with the crystals, suggesting dissolution and alteration; 6) growth of a minor amount of euhedral dolomite which was then partially transformed to calcite; and 7) neoformation of halite and pyrite.

Region B

The major differences in the mineral suite of region B relative to A are: 1) a paucity of anhydrite cement and dominance of carbonate cements; 2) more extensive and better developed quartz overgrowths; and 3) a higher proportion of detrital quartz grains. The increase in detrital quartz content continues to the north and east such that the framework grains of region C are over 95 percent quartz.

Anhydrite as cement was the first of the authigenic mineral to form, although apparently not as abundantly as farther south. Encrusting, incipient, and terminated quartz overgrowths followed on nearly all of the detrital grains. Minor amounts of authigenic feldspars, mainly K-feldspar, then formed, followed by carbonate cementation and finally pyrite formation.

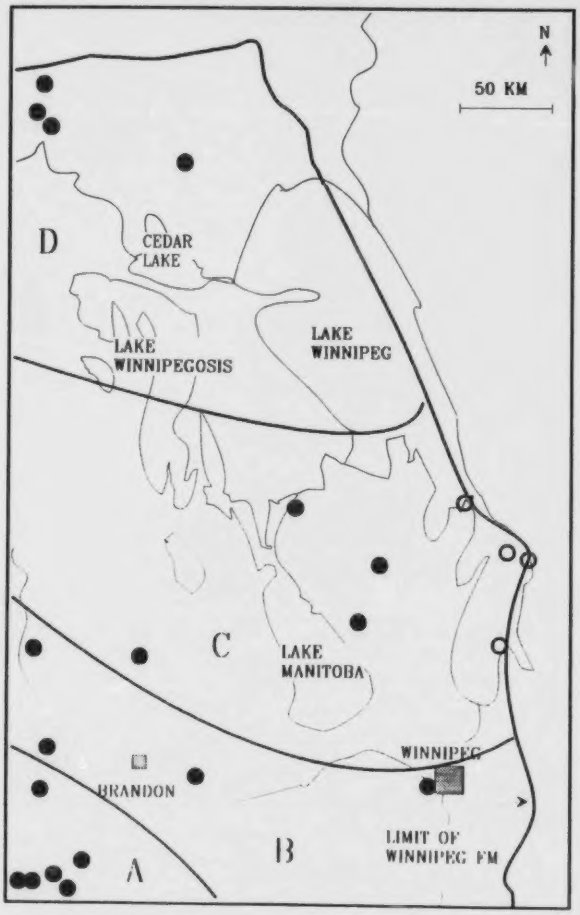

Figure 9 : Map showing the four diagenetic regions of the Black Island Member in Manitoba.

Region C

The main authigenic minerals are carbonates, clays, and quartz. Although the assemblage is relatively simple, detailed petrographic examination of the Black Island Member in this region reveals a complex sequence of precipitation and dissolution. Quartz precipitation was the first major diagenetic episode to affect these sands. K-feldspar crystallization coincided or immediately followed. These authigenic feldspars were partly transformed to illite and, more rarely, to kaolinite, although the exact timing of the transformation could not be deduced. Precipitation of pyrite in a wide variety of crystal morphologies then occurred. Finally, a second episode of K-feldspar authigenesis took place.

Region D

The northernmost region is poorly known and represented by cores from only four wells, all of which are located north of Cedar Lake. Overall, the sandstones of Region D have an authigenic mineral suite similar to C, except that illitic and carbonate cements are more common. The paragenetic sequence identified is: 1) formation of quartz overgrowths; 2) dolomite cementation;

3) dedolomitization and the formation of calcite; 4) K-feldspar precipitation; and 5) extensive illitization of the feldspars.

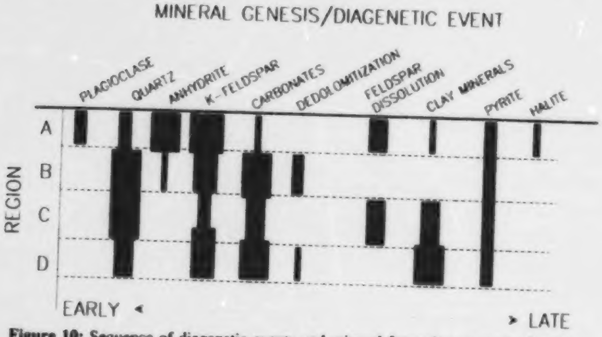

Figure 10: Sequence of diagenetic events and mineral formation recognized within each of the four major regions of the Black Island Member in Manitoba.

SUMMARY

Quartz is the dominant mineral of the Black Island Member of the Winnipeg Formation. It occurs in two forms: 1) rounded to well rounded detrital grains; and 2) incipient, encrusting, and terminated overgrowths. The Black Island samples examined contain between less than 1 percent and over 30 percent feldspar. Nearly all of this feldspar is authigenic. The feldspars are most commonly potassic and range from neoformed euhedral crystals showing very little alteration to etched and pitted crystals and detrital grains. The major clay minerals are illite and kaolinite. Illite is most common and usually shows a flaky to fibrous morphology. Kaolinite is less abundant and typically occurs as small micron-sized plates on the surface of feldspar grains. Pyrite is present in nearly all of the samples and occurs as neoformed euhedral pyritohedra, octahedra, and framboidal clusters. The most abundant cementing agents in the Black Island sandstones are anhydrite, calcite, and dolomite. Dolomite rhombs frequently show evidence of subsequent calcification. Both the anhydrite and calcite occur as pore-filling, coarsely-crystalline poikilotopic cements.

In contrast to some outcrops of the Black Island Member which contain a very simple diagenetic mineral suite, in most of the subsurface of southern Manitoba the unit has undergone a complex diagenetic history. The spatial variation of the authigenic mineral suite introduces further complexity on a regional basis as summarized in Figure 10. In general, the paragenetic sequence of the Black Island Member in southern Manitoba consists of: 1) early anhydrite cementation; 2) precipitation of quartz overgrowths; 3) formation of authigenic feldspars; 4) chemical alteration of feldspars to illite; 5) precipitation and alteration of carbonate cements; and 6) pyrite precipitation.

ACKNOWLEDGEMENTS

Our reconnaissance work on the petrography of the Winnipeg Formation could not be carried out without the assistance and cooperation of Ms. R. Bezys of the Manitoba Department of Energy and Mines. Professor E. I. Leith kindly provided samples of the Winnipeg from Manigotagan and Victoria Beach, and supplied valuable reference material. We are grateful to Drs. N. Chow and B. Sherriff of the Department of Geological Sciences, University of Manitoba, for their critical comments and many helpful suggestions.

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MINTON POOL, SOUTH-CENTRAL SASKATCHEWAN: A MODEL FOR BASEMENT INDUCED STRUCTURAL AND STRATIGRAPHIC RELATIONSHIPS

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ABSTRACT

In the Minton pool, south-central Saskatchewan, oil is trapped in structural closures in the Ordovician Red River and Devonian Winnipegosis Formation. The Minton pool marks the most northern accumulation of oil in Red River dolomites along a major oil production fairway that extends 200 km (120 miles) south into Richland County, Montana.

Development of structural closure is related to conjugate wrench faults along zones of weakness in Precambrian basement rocks. Growth of structural closures is best expressed by thinning of Silurian age strata. Upper Devonian and lower Mississippian age basement faulting is recognized along the margins of the structural closures. Porosity in both the Red River and Winnipegosis dolomites is best developed within structural closures and is commonly enhanced by fracture porosity. Likewise, porous coral-stromatoporoid bank reservoir facies of the Winnipegosis Formation crown the structural crest thereby suggesting depositional control.

Timing of structural growth, oil reserves, oil source rocks and dolomite reservoir characteristics of the Minton pool are similar to those in the well-established production trend to the south in eastern Montana. Significant exploration potential is indicated in southern Saskatchewan to at least the limits of Red River source rock maturations.

INTRODUCTION

Doming in the Minton pool area of south-central Saskatchewan has trapped oil in the Ordovician Red River C Dolomite and in the Middle Devonian Winnipegosis Formation. This oil pool situated in Township 3, Range 21W2 is the only one of its type in Saskatchewan, but is in fact the northernmost pool of a roughly north-south production trend that extends 200 km (120 miles) miles south into eastern Montana (Figs. 1,2). The Minton pool shares many of its structural, stratigraphic and reservoir characteristics with its neighbors to the south.

Estimated recoverable oil reserves per 64 ha (160 acres) in the Red River C dolomites in the Minton pool area are very near published values (225 - 340,000 bbls) for eastern Montana oil pools (Kohm and Louden 1978, 1982, 1988; Fischer *et al.*, 1990; Montana Geol. Soc., 1985). Only a fraction of the Montana Red River pools have overlying Winnipegosis oil pay; however, reserves estimates at Minton equal those in the Reserve and Wakea pools, Sheridan County, Montana (Twp 33N; Rge 56E).

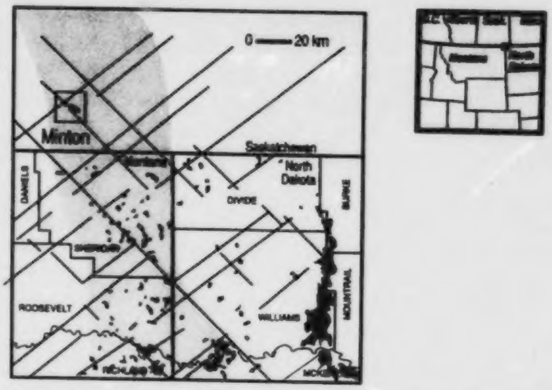


Figure 1: Regional location map of Red River oil production fairway (stipple) and Minton pool area, Saskatchewan. Conjugate basement structural lineaments adapted from Kissling & Ehrets (1988).

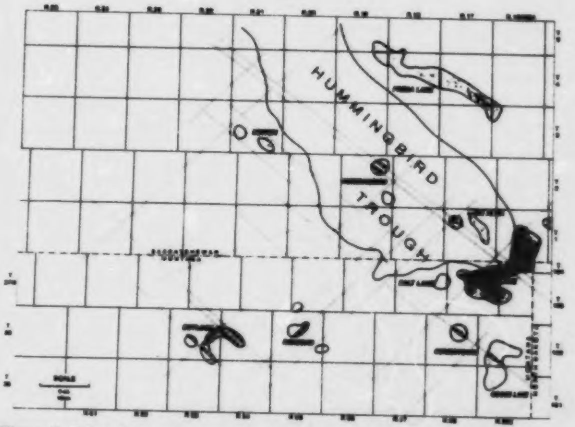


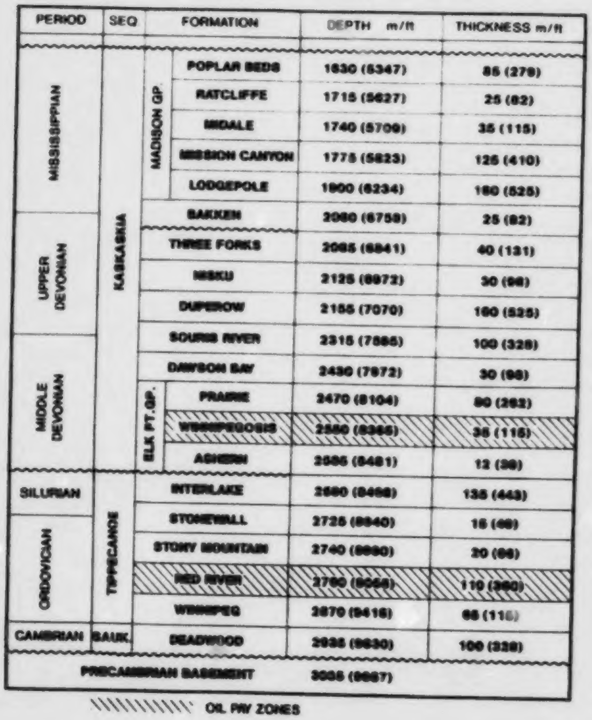
Figure 2: Oil pool location map, Minton area, Saskatchewan and Sheridan County, Montana. Basement structural lineaments shown as stippled lines.

MINTON EXPLORATION AND DEVELOPMENT

Red River oil production was first discovered in the Minton pool in July, 1976 at Dome Tenneco Minton 11-2-3-21W2 at a depth of 2,862 m (9,390 ft) [-2,088 m (-6,852 ft) subsea]. This discovery was the result of exploration by Dome Petroleum and Tenneco Oil & Gas in the mid to late 1970's and was undoubtedly prompted by exploration success for Red River C oil pools in eastern Montana. To that point only a small reserve of Red River oil had been discovered in the Lake Alma area of Saskatchewan

(Thomas, 1968)(Fig. 2). The 11-2 oil well in the Minton pool produced 20,200 m³ (127,000 bbl) over a ten year period when Mark Resources purchased the suspended well and perforated the Winnipegosis Formation at 2610 m (8,562 ft) depth [-1,835 m (-6,020 ft) subsea] and obtained an initial flow rate of 30 m³/day (190 bbl/day) at a 9/64" choke and no water cut. They drilled in December, 1989, Mark Saskoil Minton 3-17-3-21W2, 5.5 km (3.4 miles) northwest of the 11-2 recompleted oil well, and encountered 7 m (23 ft) of oil pay in the Winnipegosis and 10.5 m (34 ft) of oil pay in the Red River C dolomite. In this well, significant oil shows were also seen in the Upper Devonian Nisku and Duperow Formations as well as in the uppermost Interlake (Silurian) (Fig. 3). The 3-17 well was completed as a Red River oil well with an initial production rate of 50 m³/day (315 bbl/day) of 36°API sweet crude. The shallower Winnipegosis reservoir pay was perforated later for an initial flow rate of 80 m³/day (504 bbl/day) light crude oil (36° API).

Five additional oil wells were completed on two domal structures forming the Minton A and B pools. These are positioned along a single northwesterly trending structural nose, roughly 12 km (7 miles) long and 3 km (2 miles) in width. Cumulative Red River and Winnipegosis oil production to April 30, 1991 was 69,000 m³ (437,000 barrels).



PERIOD	SEQ	FORMATION		DEPTH m/ft	THICKNESS m/ft
MISSISSIPPIAN	KASKASKIA	MADISON GP.	POPLAR BEDS	1630 (5347)	85 (279)
			RATCLIFFE	1715 (5627)	25 (82)
			MIDALE	1740 (5709)	35 (115)
			MISSION CANYON	1775 (5823)	125 (410)
			LODGEPOLE	1900 (6234)	160 (525)
		BAKKEN		2060 (6759)	25 (82)
UPPER DEVONIAN		THREE FORKS		2085 (6841)	40 (131)
		NISKU		2125 (6972)	30 (98)
		DUPEROW		2155 (7070)	160 (525)
MIDDLE DEVONIAN		SOURIS RIVER		2315 (7595)	100 (328)
		DAWSON BAY		2430 (7972)	30 (98)
		ELK PT. GP.	PRAIRIE	2470 (8104)	80 (262)
			▨ WINNIPEGOSIS	▨ 2550 (8366)	▨ 35 (115)
			ASHERN	2585 (8481)	12 (39)
SILURIAN	TIPPECANOE	INTERLAKE		2590 (8498)	135 (443)
ORDOVICIAN		STONEWALL		2725 (8940)	15 (49)
		STONY MOUNTAIN		2740 (8990)	20 (66)
		▨ RED RIVER		▨ 2760 (9055)	▨ 110 (360)
		WINNIPEG		2870 (9416)	65 (115)
CAMBRIAN	SAUK	DEADWOOD		2935 (9630)	100 (328)
PRECAMBRIAN BASEMENT				3035 (9957)	

▨ OIL PAY ZONES

Figure 3: Paleozoic stratigraphic column, Minton pool area.

REGIONAL STRUCTURE

The Minton pool is located within the northwestern quadrant of the Williston Basin roughly 160 km (100 miles) from the basin centre which is in Williams County, North Dakota. From a depth of 3,100 m (10,200 ft) at the Minton pool, the Precambrian basement surface slopes to 4,875 m (16,000 ft) depth near the basin centre at a rate of 6 m/km (30 ft/mile).

Structural contours on the Ordovician Red River Formation portray a concentric pattern with the exception of the northerly trending Nesson anticline nearest the basin centre (Gerhard *et al.*, 1982; Lefevre *et al.*, 1987; Peterson *et al.*, 1987) (Fig. 1) and the Cedar Creek anticline trending northwesterly in southeastern Montana (Clement, 1987). In addition to these major structural elements, a dominant northwesterly and northeasterly structural lineament pattern is pervasive throughout the basin (Brown, 1978; Brown and Brown, 1987; Thomas, 1974; Gerhardt, *et al.* 1982; Mollard, 1987; Soeripto, 1986; Kissling and Ehrets, 1988).

As based upon a regional analysis of structural elements, Thomas (1974) concluded that simple-shear basement coupling along zones of basement weakness created the present structural configuration in the Williston Basin with faulting, fracturing, uplifts, and downwarps along block boundaries. Brown (1978) and Brown & Brown (1987) proposed that pure shear along a wrench fault system with both vertical and horizontal displacement creates normal faults, grabens, half-grabens and horsts along block boundaries. Along the trace of a wrench fault, Brown and Brown (1987) described rotation of basement blocks by 'scissor-type faulting' in response to adjustment of compressional stress. They also demonstrated that, with the exception of Devonian and Early Mississippian age rocks, the Paleozoic rocks have a preferential northeasterly direction of left lateral shear in response to an azimuth N 12° E of maximum horizontal compression.

Distinct lithologic changes are recognized along individual basement block boundaries. These appear as variations in Paleozoic to Mesozoic stratal thicknesses, facies, subcrop porosity and diagenetic patterns which, with structural mapping, have been used to present a more accurate paleostructural map of the trend from Sheridan County, Montana to the Minton pool area of Saskatchewan. The northeasterly and northwesterly grid is apparent (Figs. 1, 2). Superimposed upon this structural pattern are shown many of the structurally controlled Ordovician Red River and Devonian oil pools. Red River oil pools in eastern Montana are located along a north-south porosity that begins in southern Richland County, Montana and trends north to Minton, Saskatchewan (Fig. 1). The trend lies oblique to the dominant orthogonal lineament pattern but coincides with a major Precambrian crustal boundary (Camfield and Gough, 1977; Majorowicz *et al.*, 1988) that is parallel to the Nesson anticline in North Dakota. Mercury injection studies of Red River C dolomite conducted by Jennings (1987) demonstrated that the best quality dolomite reservoir is confined to this north-south trend.

STRATIGRAPHY

Paleozoic strata in the Minton area are subdivided into three stratigraphic sequences which are separated by regional unconformities: Sauk (Cambrian to Lower Ordovician); Tippecanoe (Middle Ordovician to Silurian); and Kaskaskia (Devonian to Mississippian). The Sauk sequence includes the 100 m (330 ft) thick Deadwood Formation which overlies the Precambrian basement. Lying disconformably upon the Deadwood Formation is the 60 m (200 ft) thick Ordovician Winnipeg Formation of shale and sand representative of the major marine transgression leading to deposition of the Ordovician and Silurian shallow marine carbonates (Fig. 3).

Three distinct shallowing upwards marine carbonate units are recognized in the Red River Formation, and each is capped by a thin 2 to 8 m (7 to 26 ft) anhydrite bed (Red River A, B and C). Each successive brining-upward depositional cycle thins; depositional edges of each anhydrite prograde basinwards, suggesting basinal infilling (Carrol, 1978; Kohm and Louden, 1978, Kendall; 1976).

The upper Tippecanoe sequence comprises shallow water marine carbonates and minor shales and evaporites (Stony Mountain, Stonewall and Interlake Formations, Fig. 3). The unconformity at the top of the Interlake Formation marks widespread emergence.

The Kaskaskia sequence comprises the Middle Devonian Elk Point Group, Upper Devonian carbonates, and thick Mississippian carbonates (Fig. 3). The Winnipegosis Formation marks a transgression of the Elk Point seaway from northern Alberta, diagonally across Saskatchewan and into central North Dakota (Kissling and Ehrets, 1988). The Minton pool area lies 60 km (40 miles) to the southwest of the Winnipegosis basin on a broad marine shelf platform. Regressive Elk Point seas deposited 80 m (260 ft) of Prairie salts on the shelf in the Minton area and 200 m (650 ft) in the deeper pinnacle reef basin to the east.

Upper Devonian carbonate deposition is characterized by shallow marine carbonate sequences up to 30 m (100 ft) in thickness, each usually capped by a 2 to 8 m (6 to 26 ft) anhydrite bed. The Duperow Formation, with numerous brining upward cycles from 7 to 15 m (22 to 50 ft) in thickness, best demonstrates the shallow water transgressive-regressive depositional patterns of the Upper Devonian (Wilson, 1967).

The Bakken shale and siltstone (Upper Devonian to Mississippian) represent a major transgression over the upper Devonian unconformity surface, followed by deposition of a thick overall shallowing-upwards Mississippian sequence (Lodgepole to Poplar Formations; Fig. 3). Jurassic to Cretaceous shallow marine clastic sediments lie unconformably upon the Mississippian Poplar Formation.

STRUCTURAL MODEL, MINTON POOL AREA

The two Precambrian wells at 4-32-2-20W2 and 6-11-4-21W2 demonstrate either topographic relief on the Precambrian surface or Cambrian age fault movement. Thickening basinward on an otherwise gradual slope (Carlson, 1987), the Cambrian Deadwood Formation at 6-11-4-21W2 has 21 m (70 ft) of additional sand and siltstone versus 4-32-2-20W2 (Fig. 4); presumably the additional clastics flank a paleotopographic high. Coincidentally, the thickened Deadwood at 6-11-4-21W2 lies along the eastern edge of a northeasterly trending seismic-defined low with up to 62 msec of basement relief. The relationship is comparable to the northwesterly trending graben in the Newport Deadwood sand pool in Renville County, North Dakota (T163 N, R87 W) (Clement *et al.*, 1987).

Above the Deadwood clastics to at least the Cretaceous Second White Specks Formation at roughly 960 m (3,150 ft) depth, additional structural displacement can be identified across the Minton basement structure. A Second White Specks to Winnipeg shale seismic isochron (Fig. 5) indicates 20 msec of thinning along a northwesterly trending positive high with parallel lows. Local closures are seen along the axis and may represent intersection with northeasterly trending faults. Actual amplitude on the Winnipeg shale across the Minton pool may be 30 m to 40 m (90 ft to 130 ft) (Fig. 6). Isopachs of the Cretaceous Second White Specks to the Red River C Anhydrite confirm the isochron thinning with a 37 m (121 ft) variation between 11-2 and 15-12-3-21W2 in the Minton B pool (Fig. 7).

Structure maps on the Ordovician Red River C Anhydrite, Devonian Winnipegosis and Bakken Formations, and Mississippian Greenpoint Anhydrite (Poplar Formation) all match the seismic isochron maps. On the lowermost Red River C Anhydrite (Fig. 8), up to 30 m (90 ft) of relief is seen along the basement structure. Closure diminishes to 20 m (66 ft) on the Winnipegosis Formation. Both Bakken Formation and Greenpoint Anhydrite plunge southeasterly on an anticline with 16 m (52 ft) and 11 m (36 ft) relief respectively (Figs. 9, 10, 11).

This structural relief from basement to the Second White Specks resembles those mapped in the eastern Montana Red River trend. In the Fairview pool, Richland County, Montana (Walsh, 1969) a northwesterly trending nose shows up to 70 m (225 ft) of relief on the Red River and structural closure up to at least the Bakken Formation. Above the Bakken, as at the Minton pool, shallower strata feature broad southeast-plunging anticlines. In addition to the major northwesterly trending elements, basement-related faults are also recognized on seismic lines trending northeasterly. These lower displacement faults create the up-dip structural rollover necessary to trap oil in the Red River and Winnipegosis Formations along the western edge of the Minton A pool (Fig. 5).

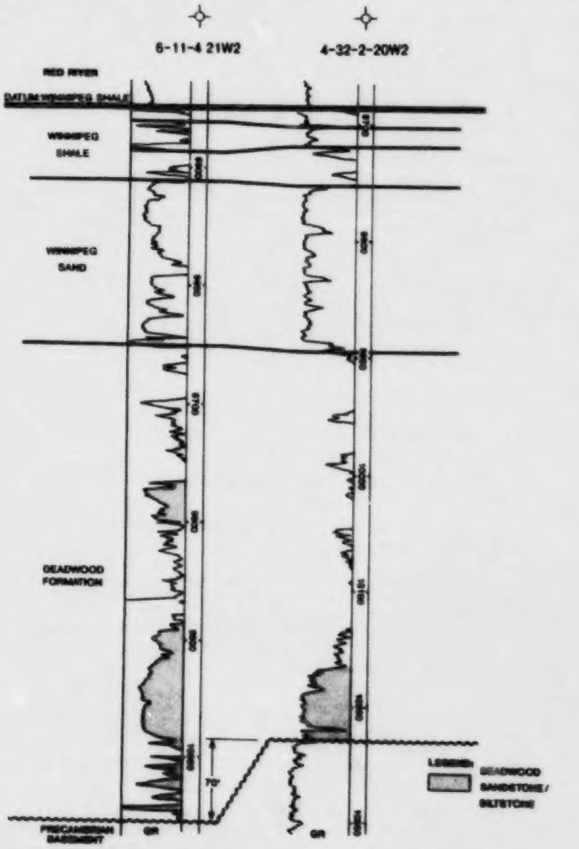


Figure 4: Cambrian Deadwood basal clastics (stippled) related to Precambrian basement relief, Minton pool area (Winnipeg shale datum).

Post-Second White Specks faulting is demonstrated on a seismic line across the Minton A pool between 3-17 and 5-17-3-21W2. A shallow thrust plane striking northwest and dipping to the southwest with up to 7 m (22 ft) of Second Specks vertical displacement is seen soling out within Gravelbourg shale (Jurassic, Fig. 12). This shallow fault presumably represents Laramide compression.

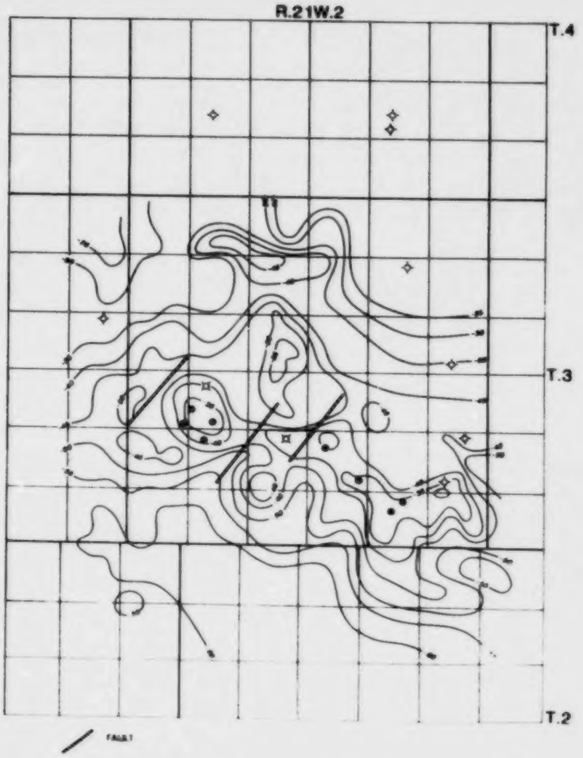


Figure 5: Second White Specks (Greenhorn) to Winnipeg shale seismic isochron, Minton pool area. Add 900 msec to map value for total isochron.

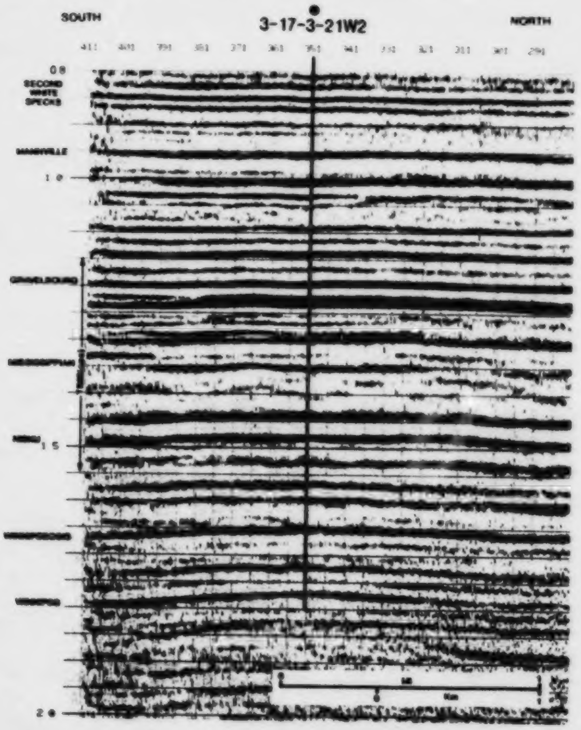


Figure 6: A north-south seismic line through 3-17-3-21W2 illustrating the Winnipegosis and Winnipeg shale relief upon an elevated basement block.

To demonstrate timing of structural movement on the Minton structure, key isopach intervals are shown graphically on Figure 7 across the Minton B pool. The Poplar Greenpoint Anhydrite (Fig. 3) is the stratigraphic datum chosen for the structural model. Between the Red River C Anhydrite and the Winnipeg Shale, only 4 m (13 ft) of thinning is seen from the structural low at 15-12-3-21W2 to the closed structural high at 11-2-3-21W2. Above the Red River C Anhydrite to the Silurian Middle Interlake marker, no isopach variations are seen relevant to the basement structure. Due to poor log quality the Middle Interlake marker at 15-12-3-21W2 is not picked.

From the Middle Interlake marker to the Devonian Ashern Formation top significant structural movement is indicated. On a percentage basis, relative structural movement between 15-12 and 11-2-3-21, measured by stratigraphic thinning of Interlake strata upon an unconformity high accounts for 28 percent of the total; 3-12-3-21W2 versus 11-2-3-21W2 is greater at 30 percent of total movement (Fig. 7). This observation is compatible with that seen in many Montana Red River oil pools (Kohm and Louden, 1988) with regional northwesterly trending nosing thins plunging to the southeast with an amplitude of 8 to 16 km (5 to 10 miles). Of particular interest is the thickness of the Devonian Ashern Formation. As in the Raymond and Outlook pools in Sheridan County, Montana (Fig. 2), thinning of the Interlake over an unconformity high associated with a positive basement structure coincides with Ashern thinning of up to 10 m (30 ft).

The Devonian Elk Point isopach (Winnipegosis Formation base to top of the First Red Beds) indicates significantly less

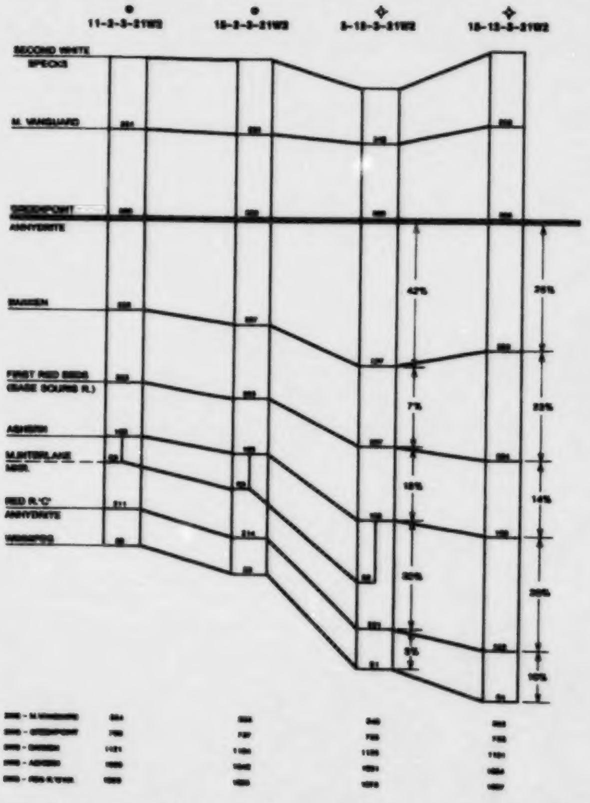


Figure 7: Isopach variations (metres) between major seismic stratigraphic markers indicating relative structural movement. Percentages next to 3-12 and 15-12-3-21W2 express temporal proportion of total Paleozoic thickening, or faulting, relative to 11-2-3-21W2 situated upon the structural crest.

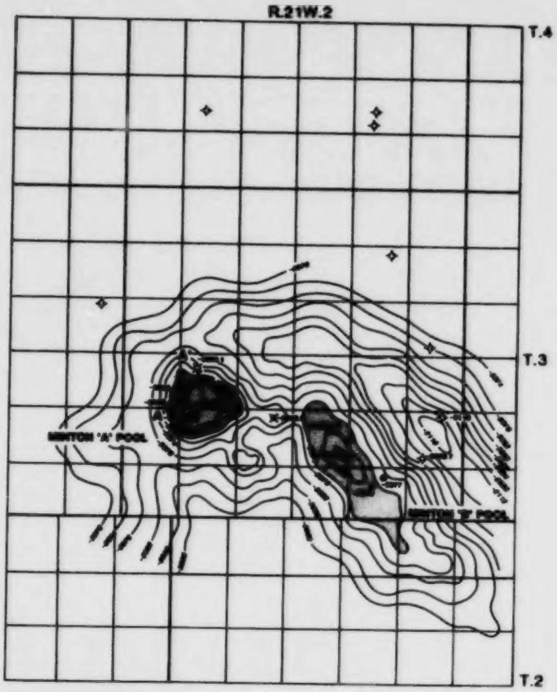


Figure 8: Structure contours on Red River C Anhydrite top, Minton pool area. Contour interval: 4 m (13 ft).

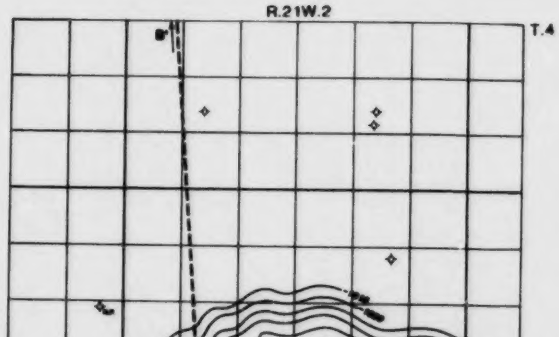


Figure 9: Structural contours on Winnipegosis Formation top, Minton pool area. Contour interval: 4 m (13 ft).

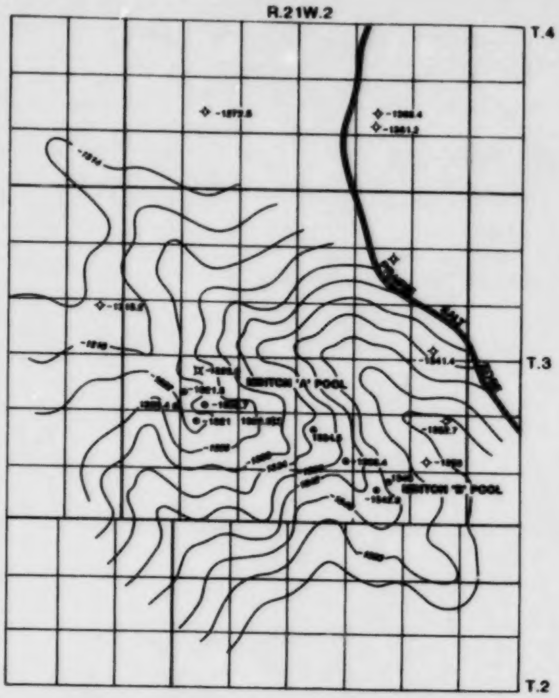


Figure 10: Structure contours on the Bakken Formation, Minton pool area. Contour interval: 4 m (13 ft).

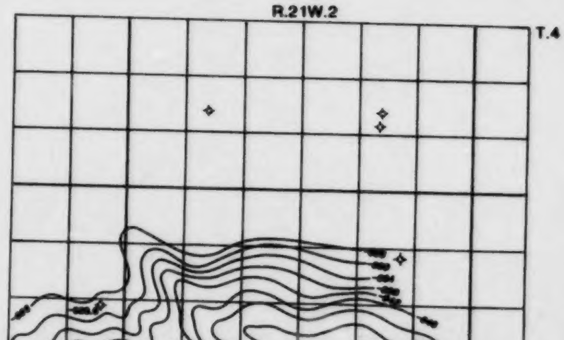


Figure 11: Structure contours on the Greenpoint Anhydrite base, Poplar Formation, Minton pool area. Contour interval: 4 m (13 ft).

structural movement, showing a range of only 6 m (20 ft), representing only 14 percent of total Paleozoic relative movement between 11-2 and 15-12-3-21 (Fig. 7). Upper Devonian strata from the Souris River to Bakken Formations thin by 9 m (30 ft) between 15-12 and 11-2-3-21 representing 23 percent of total thinning. However, at 3-12-3-21W2 on the flank of the Minton B pool these beds thicken by only 2 m, suggesting that with time, basement faulting moved outwards from the structural axis. In total, combined Devonian age structural development represented by stratigraphic thinning is 37 percent between 11-2 and 15-12-3-21W2 in the Minton B pool (Fig. 7).

Significant Mississippian structural movement as measured between the top of the Bakken Formation to the Greenpoint Anhydrite is recognized. A maximum isopach range of 14 m (46 ft) is seen primarily in the Lodgepole Formation between 3-12-3-21 and 11-2-3-21W2 representing 42 percent of total Paleozoic relative movement (Fig. 7). Basement faulting to create this broad isopach range obviously shifted back towards the structural high during Mississippian time.

The consistent thickening pattern throughout the Paleozoic section between 11-2 and 15-12-3-21W2 in the Minton B pool reverses partially in the gross interval from the Greenpoint Anhydrite (Poplar Formation) to the Cretaceous Second White Specks. Isopach thinning of the Jurassic and Cretaceous strata by up to 9 m (30 ft) at 3-12-3-21W2 on the Minton B pool flank versus 11-2-3-21W2 and 15-12-3-21W2 (Fig. 7) suggests late structural reversal, possibly related to Laramide tectonic events. This reversal accounts for the Red River C Anhydrite structural displacement of only 27 m between 11-2-3-21W2 and 3-12-3-21W2 and Paleozoic isopach thickening of 33 m (108 ft), against 38 m (125 ft) of structural relief on the C Anhydrite between 11-2-3-21W2 and 15-12-3-21W2 and 37 m of Paleozoic thickening with no Greenpoint Anhydrite to Second Specks displacement.

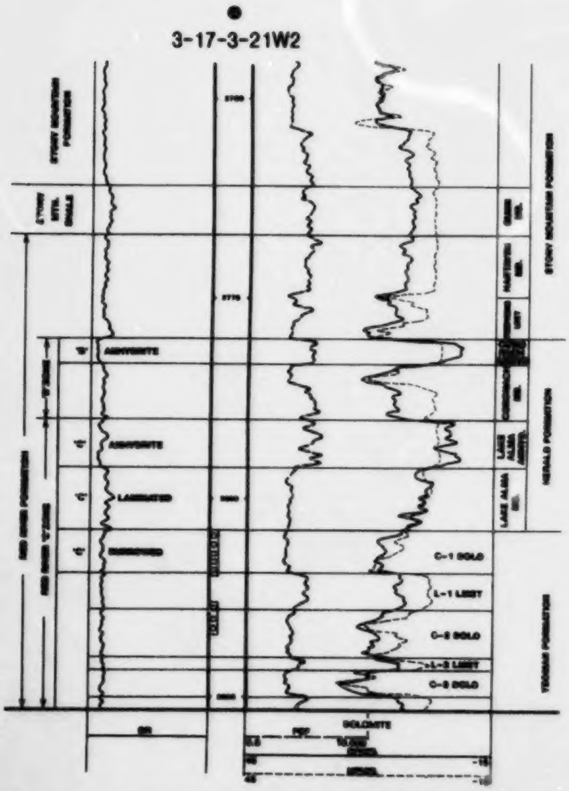


Figure 12: Red River Formation stratigraphy, Minton pool area (Mark Saskoil Minton 3-17-3-21W2)

RED RIVER RESERVOIR

Three subtidal carbonate to anhydrite cycles termed the A, B and C zones from top downwards, are encountered in the Ordovician Red River Formation (Fig. 12). Each cycle thins relative to the preceding, suggesting an overall basinal infilling.

In the lowermost Red River C cycle in the Minton pool, subtidal lime mudstone and skeletal wackestone of the C Burrowed interval (Yeoman Formation) form a lithologically homogeneous unit roughly 65 m (213 ft) in thickness. This unit features abundant dolomitized burrows which create the characteristic mottled texture (Plate 1). There is no evidence of subaerial exposure but the persistence of a limestone bed (L-1) up to 5 m (16 ft) in thickness (2,814 to 2,819 m depth) at 3-17-3-21W2 suggests early subtidal cementation. Burrows varying from 2 mm to 1 cm in diameter are well preserved in the L-1 limestone bed. Algal-rich kerogenous dolomite beds (Osadetz *et al.*, 1989) up to 20 cm in thickness are common in cores of the C Burrowed zone. The kerogenous dolomites occur either as thin (0.5 - 3 cm), dark grey to black laminated beds or as thicker (20 cm) algal-rich mudstone or wackestone. A kerogenous dolomite sample from the C Burrowed zone core at 1-10-3-21W2 contained 59 percent

Plate 1: Red River C Burrowed dolomite reservoir with abundant anastomosing burrows. C-2 dolomite at 2817.7 m depth, Mark Saskoil Minton 3-17-3-21W2.

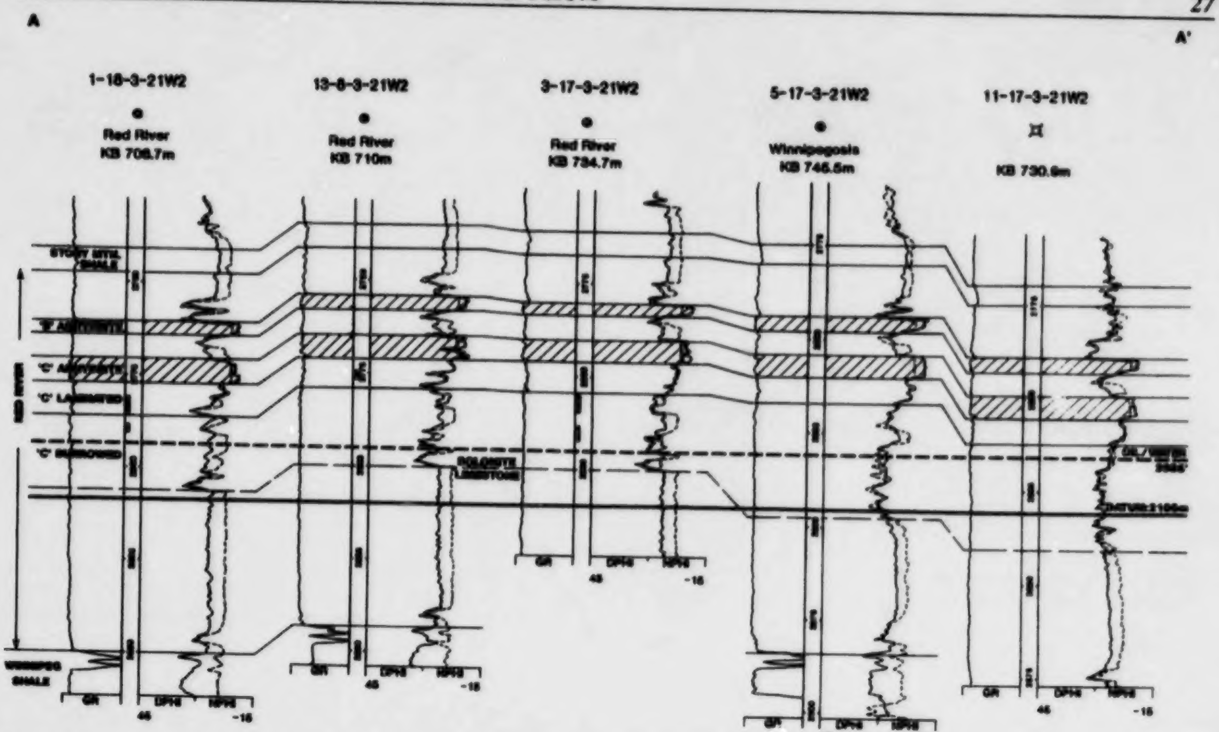


Figure 13: Structural Cross-section A-A´, Red River Formation, Minton A pool. Datum: -2100 m (6890 ft) subsea. Refer to Figure 8 for cross-section location.

T.O.C. Osadetz *et al.* (1989) noted that these kerogenous beds are also commonly recognized at contacts between the limestone beds (L-1 and L-2) and the dolomite reservoirs. Each widely distributed algal-rich bed represents temporary euxinic conditions throughout the depositional basin (Kendall, 1976).

The C Burrowed limestone and dolomite contrast sharply with the overlying 8.5 m (28 ft) thick C Laminated dolomite mudstones (Lake Alma) in the Minton pool area. A 10 to 25 cm (4 to 10 inches) interbedded zone between the two stratigraphic units suggests a continuous transition towards shallow marine conditions. The top of the Red River C Cycle is marked by the 6 m (20 ft) thick C Anhydrite (Lake Alma anhydrite). The basinwide distribution of the A, B and C anhydrites, with each anhydrite bed thickness proportional to the overall cycle thickness, supports an intrabasinal subtidal versus supratidal depositional pattern. The Red River C Anhydrite acts as an effective cap seal to Red River oil reservoirs at Minton and in Montana.

The Minton pool comprises three distinct Red River dolomite reservoirs ranging from 3 to 10 m (10 to 33 ft) in thickness (Fig. 12). Depth of dolomitization into the Red River C Burrowed zone beneath the Red River C Anhydrite is 29 m (95 ft) in most wells; a maximum depth of 39 m (128 ft) is seen at 5-17-3-21W2 and 11-2-3-21W2, both near the centre of the Minton A and B pools (Fig. 13). The uppermost C-1 dolomite reservoir lies between the C Anhydrite base and the L-1 limestone bed top (Fig. 12). In this upper reservoir zone, greatest intercrystalline dolomite porosity occurs immediately above the L-1 limestone bed in the upper 5 to 6 m (10 to 20 ft) of the C Burrowed zone (10 - 13 percent; 5 - 231 mD; 3-17-3-21, Plates 2, 3). Porosity decreases to zero upwards through the C Laminated dolomite, but in which subvertical open fractures are common. Greatest C Laminated porosity coincidentally occurs nearest basement structure edges in both the Minton A and B pools (13-8 and 1-18-3-21; 15-2 and 3-12-3-21; Fig. 13). In spite of a 5 to 8 m (16 to 26 ft) Red River structural decline between 3-17-3-21 and 1-18-3-21 in the Minton A pool, structure on the porosity top differs by only 1.5 m (5 ft) leaving approximately equivalent net pay values. Across the Minton A and B pools, the L-1 limestone serves to isolate the C-1 and C-2 dolomites. The oil-water contact of the C-1 dolomite in the Minton A pool lies at approximately -2,085 m (-6,840 ft) subsea. In the Minton B pool, along the basement structural edge, the oil-water contact at 15-2-3-21W2 is down along the flank of the structure at roughly -2,098 m (-6,883 ft) subsea.

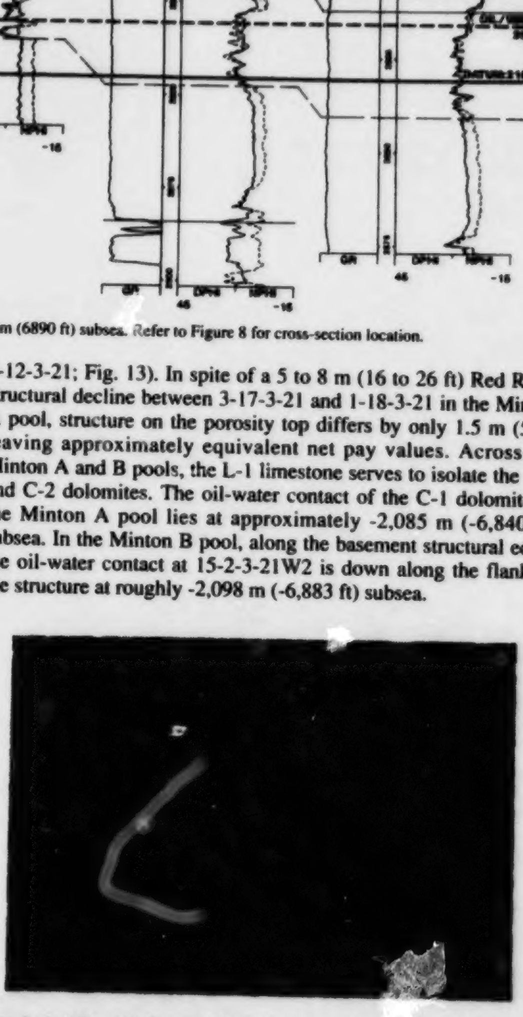


Plate 2: Red River C Burrowed dolospar porosity. C-2 dolomite reservoir at 2817.7 m depth, Mark Saskoil Minton 3-17-3-21W2. PPL, 100X.

The C-2 dolomite reservoir lies between the L-1 and L-2 limestone beds (Fig. 12). Thickness of the interval ranges from 6 m (20 ft) at 3-17 and 13-8-3-21W2 to 10 m (32 ft) at 5-17-3-21W2 on the Minton A pool dependent upon thinning of the overlying L-1 limestone bed. In fact, at 11-17-3-21W2, the L-1

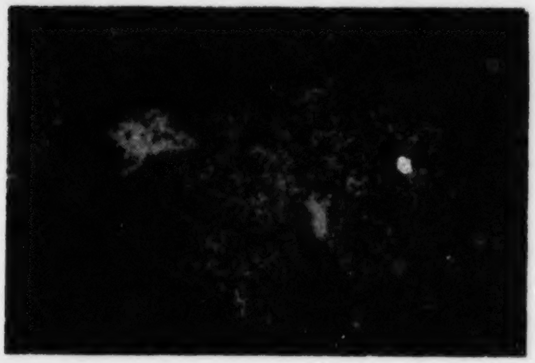


Plate 3: Red River C Burrowed dolomite. Nodular texture defined by dark anhedral dolomicrospar on burrow boundaries surrounding intercrystalline/ microvugular porosity. C-2 dolomite at 2858.6 m depth, Mark Saskoil Minton 1-10-3-21W2. Photo from Jackalope Geological Report 9OJGO8. PPL, 25X.

and L-2 limestone beds are both absent, thereby creating a single dolomite with an average of 12 percent porosity. Maximum porosity in the C-2 dolomite varies from 12 to 18 percent with a permeability range of 2 to 82 mD, and consistently is highest beneath the thick 5 m (16 ft) L-1 limestone bed (3-17 and 5-17-3-21 W2, Fig. 13). The oil-water contact in the C-2 dolomite lies at roughly -2,085 m (-6,840 ft) subsea in the Minton A pool (Fig. 13); maximum net pay of the C-2 dolomite is 5 m (16 ft) at an 8 percent cutoff (3-17-3-21W2). The C-2 reservoir tested wet at 1-10 and 11-2-3-21W2 in the Minton B pool.

The lowermost C-3 dolomite reservoir is in most cases up to 3.5 m (12 ft) thick and in sharp contact with the lower C Burrowed limestone. Exceptions to this reservoir configuration are seen in the Minton A pool at 5-17 and 11-17-3-21W2 and in the Minton B pool at 11-2-3-21W2 where the C-3 reservoir zone is up to 9 m (30 ft) thicker. Rather than a sharp lower contact, the base of the C-3 reservoir zone in such wells is typically more transitional over 5 to 8 m (16 to 26 ft) into the underlying limestones. Porosity and permeability are greatest where the C-3 reservoir is in sharp contact with underlying limestone, as seen in the C-1 reservoir. The C-3 dolomite reservoir is always wet in both the Minton A and B pools.

Red River C porosity development across the Minton A and B pools is greatest along the axis of the structural high (5-17-3-21W2) and coincides with maximum vertical depth of dolomitization (39.2 m, 129 ft) at 5-17-3-21 versus 28.7 m (94 ft) at 1-18-3-21W2 (Fig. 13). Depth to first porosity beneath the Red River C Anhydrite is greatest at 11-17-3-21 on the structural flank (13.7 m, 45 ft), decreasing to 5 to 10 m (16 to 32 ft) upon the structural crest where greatest net porosity is encountered at 3-17 and 5-17-3-21W2 (22 m and 19 m respectively). Porosity variability is seen primarily within the C-1 and C-2 reservoir zones.

Processes of dolomitization and porosity development of the Red River C Laminated and Burrowed beds have been discussed in detail by Kendall (1976), Kohm and Louden (1978, 1987), Longman *et al.* (1983) and Asquith *et al.* (1978). Each author proposes post-depositional gravity driven migration of magnesium-rich brines into underlying Red River limestones. Kendall (1976, 1977, 1979) proposed Devonian Prairie brine migration as partially responsible for porosity in Red River dolomite. The Richland County dolomitization model, proposed by Kohm and Louden (1978), involves migration of Silurian age brines along dominant northeast-southwest trending basement fracture zones, creating narrow 1 to 2.5 km wide (0.6 to 1.5 miles) northeasterly porosity trends. Longman (1982) and Longman *et al.* (1983) proposed magnesium-rich brine migration downward into Red River C limestones during precipitation of the Red River C Anhydrite bed. This model assumes fluid transmission through holes in the gypsum bed, possibly associated with an underlying basement structure. Depth of fluid penetration in the creation of dolomite reservoirs would be dependent upon a number of factors including: i) hydrodynamic head of the dense brines versus the Red River formation waters; ii) high lateral transmissibility of thin beds in the C Burrowed to inhibit downward movement; iii) presence of low permeability algal kerogenite beds as a barrier to vertical fluid migration, and iv) vertical fractures to move fluids deeper into the Red River C Burrowed zone.

Although reservoir variability in the Red River C zone across the Minton A and B pools has much in common with that described by Longman *et al.*(1983) and Kohm and Louden (1978, 1988), the following are noted:

1) Three gross dolomite reservoir beds separated by laterally persistent limestone beds are traced with remarkable continuity across and down the axis of the basement structure and beyond for at least 8 km (5 miles) from the Minton pools to 7-32-2-22W2 and 15-26-3-21W2.
2) Maximum depth of dolomitization beneath the C Anhydrite varies by a maximum of 10 m (33 ft), regardless of structural orientation. The Richland County Model demonstrates wide variation in dolomitization patterns that depends upon position relative to structural closures on larger structural crests.
3) The Richland County pools include a tight zone in the Red River C Laminated and Burrowed unit of cryptodolomite (Longman *et al.*, 1983) surrounded by microsucrosic dolomite reservoirs. This pattern is not recognized in the Minton pool.

WINNIPEGOSIS RESERVOIR

At the southwestern end of the Middle Devonian Elk Point basin centred in southeastern Saskatchewan and north-central North Dakota, a shallow marine Winnipegosis platform surrounds the deeper basin which is characterized by pinnacle reef growth. The Minton pool lies on the western platform, roughly 58 km (36 miles) southwest of the northwesterly trending shelf margin.

Winnipegosis shelf facies prograde across the western platform to the shelf margin (Kinard *et al.*, 1978; Kissling and Ehrets, 1988) resulting in formation thickening from 26 m (84 ft) at the Outlook pool, Sheridan County, Montana, to 34 m (112 ft) at the Minton pool to at least 56 m (185 ft) at platform reefs near the shelf margin at Tatagwa, Saskatchewan (Township 6; Range 15W2M) and at the Temple pool, Williams County, North Dakota (Kissling and Ehrets, 1988). In general, the presence of progradational *Amphipora* shoals and coral-stromatoporoid banks eastward towards the shelf margin is accompanied by increasing thickness and net porosity. Structural movement of basement blocks along northwesterly and northeasterly oriented lineaments created complex depositional sub-basins on the Winnipegosis platform.

In the Minton area the Winnipegosis shelf carbonates are subdivided into three overall shallowing upwards sub-units from base to top: 1) Marine shelf facies, 2) Coral-stromatoporoid and *Amphipora* shoal facies and 3) the peritidal facies (Fig. 14).

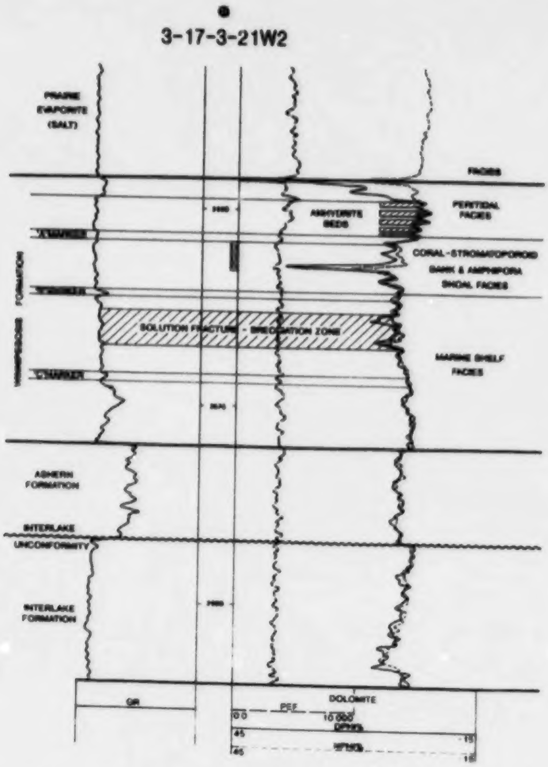


Figure 14: Winnipegosis Formation stratigraphy, Minton pool area (Mark Saskoil Minton 3-17-3-21W2).

MARINE SHELF FACIES

The basal Winnipegosis marine shelf facies is up to 20 m (66 ft) thick in the Minton pool (Fig. 14). The unit beneath the C argillaceous marker is up to 9 m (30 ft) thick and is a tight nodular mudstone (dolomicrospar). Between the C and B argillaceous markers are mixed nodular-bedded mudstones and skeletal wackestones, commonly with abundant white anhydrite. In addition to intercrystalline (dolospar)/microvugular porosity in the wackestone, fracturing and solution-collapse brecciation add to overall reservoir porosity (up to 18 percent, average 6.5 percent) and permeability (greater than 2 D) (Plates 4, 5). Fracture porosity is estimated to contribute up to 4 percent of the total porosity. Dissolution along subvertical fractures created large (up to 2 cm wide) irregularly shaped vugs; commonly an earthy light-brown dolomite rim is seen along vug boundaries. Best-developed dolospar and fracture porosity is encountered on the axial high, particularly in the Minton A pool (3-17 and 5-17-3-21W2; Plate 5). Absence of white anhydrite infill of vuggy fractures (Plate 5)upon the axial high suggests late dissolution, possibly synchronous with that of the total Prairie salt in the Hummingbird Trough immediately to the east (Fig. 2).

AMPHIPORA SHOAL AND CORAL-STROMATOPOROID BANK FACIES

This facies, with up to 7 m (23 ft) of oil pay (3-17-3-21W2; Figs. 14, 15), lies between the A and B argillaceous markers. Up to 5 m (16 ft) of intercrystalline dolospar with porosity greater than 7 percent is found; permeability ranges from 2.5 to 190 mD (Plate 6). In Township 4, Range 21W2 to the north of the Minton

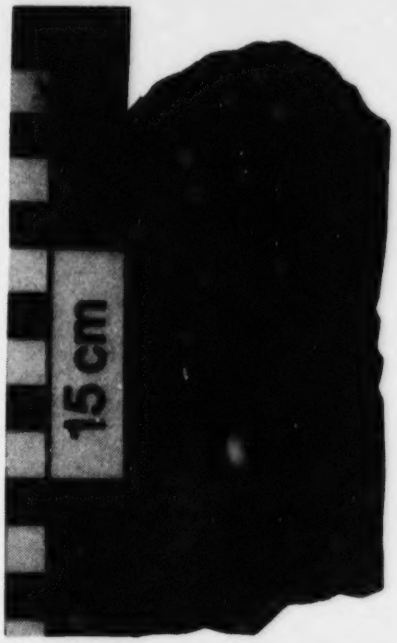


Plate 4: Fracture porosity, *Amphipora*-skeletal wackestone, marine shelf facies, Winnipegosis Formation. Mark Saskoil Minton 3-17-3-21W2; core depth 2560.5 m.

Plate 5: Solution collapse breccia within nodular bedded skeletal wackestone marine shelf facies, Winnipegosis Formation. Mark Saskoil Minton 3-17-3-21W2; core depth 2565 m.

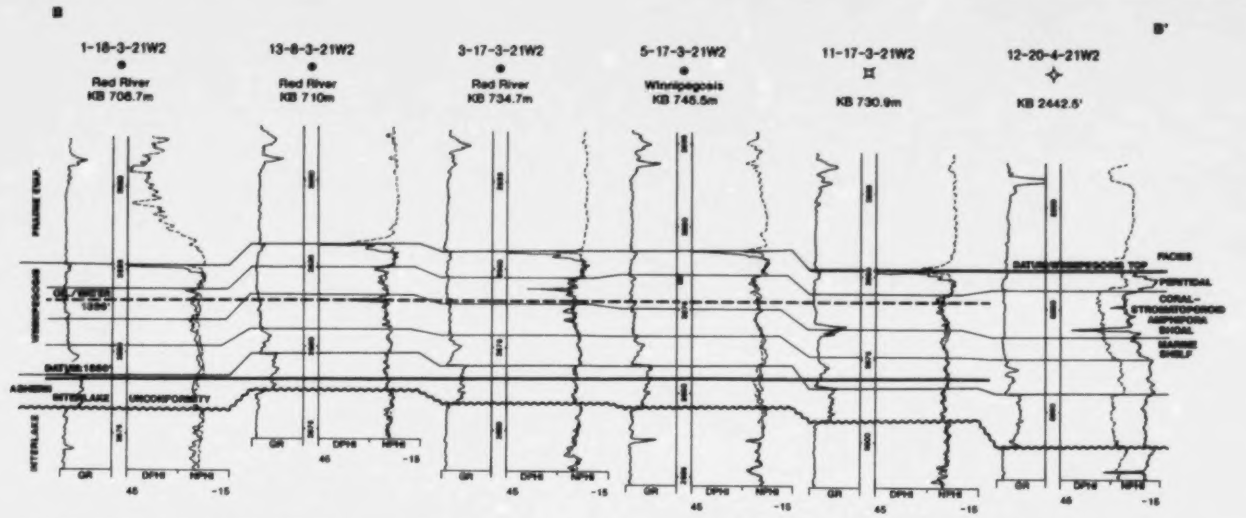


Figure 15: Structural cross-section B-B′, Winnipegosis Formation, Minton A pool. Datum: -1850 m (6070 ft) subsea. Refer to Figure 9 for cross-section location.

pools, three Winnipegosis wells encountered a thicker middle porous zone, primarily at the expense of the upper and lower (Fig. 15). None of these wells were cored, but log porosity in the mid-Winnipegosis zone at 12-20-4-21W2 is up to 16 percent over a maximum thickness of 12 m (40 ft) (Fig. 15).

Major constituents include coral-stromatoporoid floatstone, *Amphipora*-peloidal wackestone to packstone, and lesser skeletal wackestone to mudstone. *Amphipora* plus laminar and tabular stromatoporoids, crinoids and brachiopods are major skeletal components in this facies. *Thamnopora*, *Alveolites*, *Stachyodes*, globular stromatoporoids and *Renalcis* algae are less abundant. The coral-stromatoporoid floatstone is most common near the top of this unit and occurs in thicknesses up to 4 m (13 ft) at 1-10-3-21W2. The floatstone unit in 3-17-3-21 (2,554-56 m) represents the best Winnipegosis reservoir with a maximum 10 percent porosity and 190 mD permeability. Sub-vertical open fractures are common throughout. In contrast, at 15-2-3-21W2 less than 1 m (3 ft) of floatstone with less than 6 percent porosity and less than 2 mD permeability is found on the flank of the structure. As in the Outlook and Raymond pools, Sheridan County, North Dakota (Parker *et al.*, 1982) significant oil reserves are recovered from tight dolomicrospar mudstones to packstones that have been intensely fractured upon the crest of a basement structure (Plate 5). Such fracture porosity is believed to contribute significantly to overall reservoir quality at 11-2-3-21W2.

PERITIDAL FACIES

The uppermost Winnipegosis peritidal facies is a very dense laminated mudstone (dolomicrospar) with rare peloids and stromatolites (Fig. 14). Near the base are thin interbeds (0.25 m) of peloidal-skeletal wackestones to packstones. The uppermost 2.5 m (8 ft) of the facies are more argillaceous and include abundant horizontal displacive halite separations. Interbedded with the dense laminated mudstones in this peritidal facies zone are some five distinct nodular-mosaic to laminated anhydrite beds, each up to 0.25 m thick (Fig. 14). The peritidal facies in the Minton pools is up to 7.5 m (25 ft) thick (3-17-3-21W2) and includes five anhydrite beds versus 4.8 m (16 ft) thickness and two anhydrite beds immediately to the north at 12-20-4-21W2 (Fig. 15). Thinning of the peritidal facies is associated with thicker porous *Amphipora* shoals/coral-stromatoporoid banks immediately beneath as seen at 12-20-4-21W2. This means that in both the Minton A and B pools, Winnipegosis reservoirs are thinner and less porous than those found less than 10 km (6 miles) north in the Township 4, Range 21 wells. This relationship is seen continuously along the western platform to at least the Reserve and Wakea oil pools in Township 33N, Ranges 55 and 56E, Sheridan County, Montana.

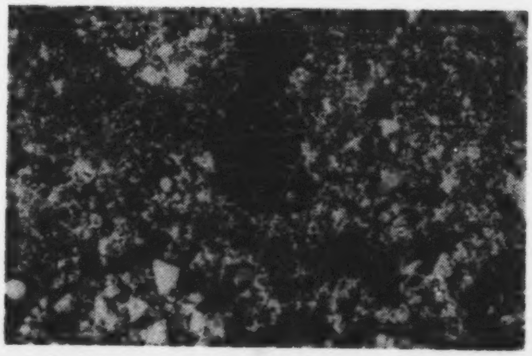


Plate 6: Coral stromatoporoid floatstone, Winnipegosis Formation. Original depositional textures replaced by coarse dolospar, vugs and intraskeletal voids. Note open fracture. Mark Saskoil Minton 15-9-3-21W2; core depth 2545.3 m. PPL, 25X.

Oil-water contacts in the Minton A and B pools lie at roughly -1,826 m (-5,991 ft) and -1,842 m (-6,043 ft) respectively. A strong water drive is provided through the fracture porosity in the underlying marine shelf facies.

OTHER PROSPECTIVE HORIZONS

Oil shows in dolomites of the Upper Devonian Duperow and Nisku Formations (Fig. 3) are common in both Minton oil pools. The Duperow Formation is characterized by numerous cyclical progradational carbonate sequences of lime mudstone to packstones plus thin stromatoporoid banks, capped by thin 2 to 5 m (7 to 16 ft) anhydrite beds (Wilson, 1967; Dunn, 1975; Kissling and Ehrets, 1988). The best dolospar porosity occurs in thin beds (2 to 5 m) (7 to 16 ft) immediately beneath each anhydrite. In the Minton pool, individual dolomite reservoirs with good oil shows can be traced for over 30 km (20 miles) south to the Raymond pool, Sheridan County, Montana (Parker *et al.*,

1982) (Fig. 2) which has oil reserves of up to 64,000 m^3 (400,000 barrels) per 160 acres. Fresh water flushing of Upper Devonian strata (McTavish and Vigrass, 1987) may explain the lack of Duperow oil production in the Minton pool.

In the Nisku Formation up to 8 m (20 ft) of dolomitized lime mudstone to *Amphipora*-rich wackestone to packstone lie beneath an anhydrite up to 9.5 m (31 ft) thick. Whereas light oil was recovered in a drillstem test at 15-2-3-21W2, there has been no commercial discovery. The Hummingbird pool with Nisku structural closure upon a multi-stage Prairie salt solution feature lies just 20 km (12 miles) to the east.

Abundant degraded oil shows in the Mississippian Mission Canyon and Ratcliffe Formations are found in both the Minton A and B pools. Meteoric water flushing of Mississippian strata is common in the Minton area, eliminating production potential.

SOURCE ROCKS

Algal kerogenites of the Red River Formation, composed primarily of the alga *Gloeocapsomorpha prisca* (Foster *et al.*, 1989; Osadetz *et al.* 1989), are considered the source of Red River oil in the Minton pool. A regional ROCK-EVAL pyrolysis evaluation of 66 Red River C kerogenite samples by Osadetz *et al.* (1989) identified potential Type I source rocks with an average TOC (total organic carbon) of 8.4 percent and an average yield of 71.7 kg (158 pounds) hydrocarbon per tonne of rock (S2). The nearest sample to the Minton pool at 6-11-4-21W2 contained 6.6 percent TOC and a hydrocarbon yield of 61.4 kg/tonne. Depth of thermal maturity is estimated to lie at roughly 2,950 m (9,678 ft) (Ozadetz *et al.*, 1989). The Red River C kerogenites lie at from 2,825 m (9,268 ft) to 2,875 m (9,432 ft) depth in the Minton pool. Abundant light oil was also encountered in Red River core at Mark Bengough 7-10-4-24W2 at a depth of 2,640 m (8,661 ft) and is suggestive of oil migration beyond limits of thermal maturity.

An analysis of Winnipegosis oil from 11-2-3-21W2 in the Minton pool confirms a Red River kerogenite source, with migration to the reservoir along vertical fractures (Osadetz, pers. comm.). This conclusion is compatible with an oil correlation study in the Raymond pool, Sheridan County, Montana (Fig. 2) that demonstrated a common Red River source for both the Winnipegosis and Red River oil samples (Leebeer and Zumberge, 1987).

SUMMARY

Structural-stratigraphic relationships relevant to oil entrapment in the Minton oil pool are summarized as follows:

1) Wrench faulting of Precambrian basement rocks associated with regional compressive forces created an elevated basement block, that based upon seismic data, is roughly 3 to 12 km (2 by 7 mi) in dimension and trends northwest-southeast. Maximum vertical relief on the Precambrian basement surface is 30 to 40 m (100 to 130 ft). Intersection of the basement block with secondary northeast-southwest trending basement faults has created at lease three isolated structural closures up to 320 ha (800 acres) in size.

2) Paleoerosional relief on the Precambrian basement is commonly recognized on seismic lines by presence of thickened Cambrian Deadwood clastics in basement structural lows. Timing of greatest structural movement on basement is defined by erosional thinning of upper Interlake strata and is related to major vertical movement. Relative structural movement across the Minton basement block is recognized sporadically in Upper Devonian strata and more significantly in lower Mississippian Lodgepole Formation time.

3) Thickest coral-stromatoporoid bank and *Amphipora* shoal reservoir facies of the Winnipegosis Formation correspond to the Minton basement structure crest; no depositional influence relative to the basement structure is seen in the Red River Formation.

4) Greatest Winnipegosis and Red River secondary dolomite porosity is related to structural closure, particularly in the Minton A pool located farthest up-dip in the basement block. Dolospar is most common in Winnipegosis dolomite upon structural closures versus dolomicrospar nearest the edge of the basement block. Dissolution and collapse brecciation of the middle Winnipegosis marine shelf facies are most common at the structural crest (3-17-3-21W2); anhydrite infilling of fracture-dissolution porosity is pervasive with greatest subsequent dissolution nearest structural crests.

5) Structural and stratigraphic relationships discussed can be applied to regional exploration for Minton-type oil pools in southern Saskatchewan, particularly with the application of seismic stratigraphic analysis.

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DOLOMITIZATION IN MIDDLE DEVONIAN PLATFORM CARBONATES, ELM POINT FORMATION, INTERLAKE AREA, MANITOBA

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ABSTRACT

Partially dolomitized, fine-grained limestones make up most of the Middle Devonian (Eifelian) Elm Point Formation in the Paleozoic outcrop belt in southern Manitoba. Four lithofacies are recognized: 1) nodular limestone, composed of grey-brown, skeletal wackestone-packstone nodules in a tan-weathered matrix of skeletal wackestone-packstone; 2) mottled, skeletal packstone; 3) tabular stromatoporoid-coral boundstone, and 4) skeletal-intraclast rudstone-floatstone. These lithofacies lack evidence for subaerial exposure and contain abundant lime mud and a diverse fossil assemblage, suggesting that they were deposited in a low-energy, open subtidal setting which was subjected to intermittent storms.

The extent of dolomitization in the Elm Point platform was significantly controlled by sediment permeability and porosity. Burrowing, boring and microbial micritization enhanced permeability and porosity which, in turn, locally promoted later dolomitization. In contrast, early limestone lithification, evidenced by intraclasts and hardgrounds, effectively reduced permeability and porosity, inhibiting dolomitization.

INTRODUCTION

Nodular and mottled limestones are ubiquitous components of many Paleozoic platform sequences. Characteristic textures of these limestones have been commonly attributed to bioturbation and preferential dolomitization of burrow structures. Detailed field, drill-core and petrographic studies of nodular and mottled limestones of Middle Devonian (Eifelian) age in southern Manitoba support this attribution, but indicate that other factors also helped control dolomitization in these rocks.

The purpose of this paper is to document the occurrence of dolomite in platform limestones of the Middle Devonian Elm Point Formation in selected locations in the Manitoba outcrop belt, and to demonstrate that products of early diagenesis locally controlled the extent of dolomitization. Data for this study were collected from quarries and from cores drilled near the Lake Manitoba Narrows and the adjacent area.

GEOLOGIC SETTING

The Elm Point Formation (Baillie 1953, 1955) is composed of partially dolomitized limestones in the northwest-trending Paleozoic outcrop belt in the Interlake area, southern Manitoba, between Oak Point and Waterhen Lake (Fig. 1). These limestones, 5 to 20 m in thickness, are exposed along the shores of Lake Manitoba and in numerous quarries in the area. The Elm Point Formation has also been cored by Manitoba Energy and Mines as part of a stratigraphic drilling program in the outcrop belt.

The Elm Point Formation represents shallow-platform deposition in the eastern part of the intracratonic Elk Point Basin (Grayston *et al.* 1964; Fig. 2). It is one of the major occurrences of limestone in the basin. The formation overlies argillaceous dolostone and shale of the Ashern Formation and underlies dolostone of the reef-bearing Winnipegosis Formation (Fig. 3). The Prairie Evaporite caps the Winnipegosis Formation but is not preserved in the outcrop belt. In the northern end of the outcrop belt and westward into the subsurface, Elm Point limestones grade laterally into dolostones of the Lower Member of the Winnipegosis Formation. These four formations comprise the Elk Point Group, the lowermost and best defined of the three carbonate-evaporite cycles in the Devonian sequence (Norris *et al.*, 1982).

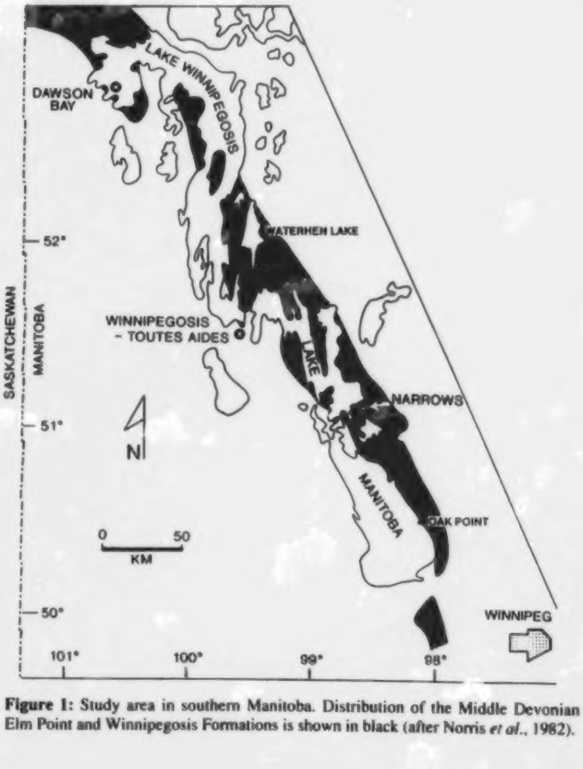

Figure 1: Study area in southern Manitoba. Distribution of the Middle Devonian Elm Point and Winnipegosis Formations is shown in black (after Norris *et al.*, 1982).

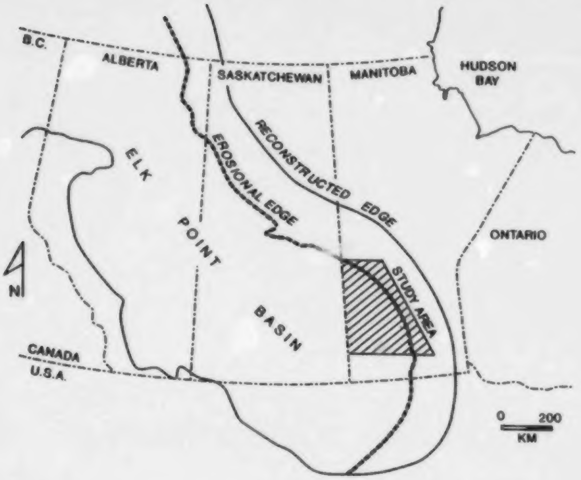

Figure 2: Geologic setting for the Elm Point Formation in western Canada (after Grayston *et al.* 1964). The study area (Fig. 1) is marked by diagonal lines.

DEPOSITIONAL FRAMEWORK

Four limestone lithofacies are recognized in outcrops and cores of the Elm Point Formation: 1) nodular limestone; 2) mottled limestone; 3) stromatoporoid-coral boundstone, and 4) intraclast floatstone-rudstone. The first three probably represent deposition in a generally low-energy, open subtidal setting. This is indicated by the lack of subaerial exposure features, by the abundance of lime mud and by the diverse assemblage of whole and comminuted skeletal components (such as brachiopods, echinoderms, bryozoans, gastropods, branching and tabular corals, stromatoporoids and calcified algae). The last lithofacies likely represents storm deposition.

The following descriptions employ Dunham's limestone classification modified by Embry and Klovan (1971).

1) Thick-bedded **nodular limestone** is the most distinctive lithofacies of the Elm Point Formation in the study area (Fig. 4a). The limestone is composed of centimetre-scale nodules of greyish-brown coloured, skeletal wackestone-packstone in a tan-weathered, internodular matrix of partly dolomitized, skeletal wackestone-packstone. The nodules are irregular in size and shape, and make up 30 to 60 percent by volume of the rock unit (visual estimates). Boundaries between nodules and matrix are sharp to gradational over millimetre distances. Horizontal and subvertical burrows are recognizable and are commonly dolomitized. Wispy, clay- and organic-rich laminae are locally developed and typically warped around allochems. Calcite-filled vugs, some of which are pseudomorphs after halite hopper crystals, are ubiquitous. V-shaped, calcite-filled fractures occur only in limestone nodules.

(2) **Mottled limestone** consists of thick-bedded, mottled, buff- and brown-coloured, skeletal wackestone-packstone. Tan-weathered, dolomitized burrows are common (Fig. 4b). Rare beds of mottled limestone with erosional top surfaces were observed in a few quarries. They probably represent submarine hardgrounds. Recessively weathered, argillaceous limestone laminae, which can be traced over the extent of the outcrop, cap some mottled limestone beds. Mottled limestones pass laterally and vertically into nodular limestones.

(3) **Stromatoporoid-coral boundstones**, up to 30 cm in thickness and several metres in lateral extent, are interbedded with nodular and mottled limestones. Tabular stromatoporoids and tabulate corals are the main constituents. They are locally associated with bulbous stromatoporoids. All these organisms are typically preserved *in situ* and, commonly, are bored. The matrix consists of partially dolomitized, skeletal packstone.

(4) **Intraclast floatstones-rudstones**, up to 5 cm in thickness, randomly punctuate the Elm Point Formation. In quarries, individual beds may be traced across the entire exposure or may be laterally discontinuous, grading into nodular limestone. Pebble-sized intraclasts are subequant, angular to rounded, and composed of lime mudstone to skeletal wackestone. The interparticle matrix is skeletal wackestone, which is locally dolomitized, and rarely skeletal grainstone.

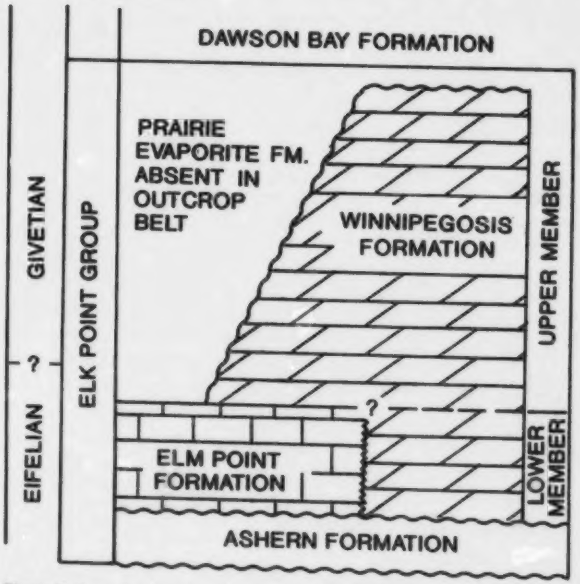

Figure 3: Generalized stratigraphic section of the Elm Point Formation and adjacent units (after Norris *et al.* 1982).

CALCITE PETROGRAPHY

Elm Point limestones, now composed of low-Mg calcite (and herein referred to as calcite), contain abundant sand-sized bioclasts (Fig. 5a, b). The most common large bioclasts are echinoderms, articulate brachiopods, stromatoporoids and bryozoans which have well-preserved internal structures. In contrast, molluscan remains are now composed largely of fine- to medium-crystalline calcite. Calcified algae include *Girvanella*, *Sphaerocodium* and *Wetheredella*, as well as poorly preserved remnants of probable codiacean algae. *Girvanella* and *Sphaerocodium* commonly occur in sand-sized particles of intertwined tubules (Fig. 5a) and encrusting stromatoporoids and corals. Sand- and granule-sized intraclasts of skeletal wackestone-packstone are scattered throughout the beds and contain skeletal particles which are truncated by intraclast margins.

Many bioclasts are enveloped by micrite up to 0.5 mm in thickness. The envelopes preserve the original outline of altered particles and, in some, tubules of *Girvanella* are recognizable.

Micrite-filled borings are also common in bioclasts and are typically irregular or vermiform in shape with a diameter less than 0.5 mm (Fig. 5b). The smaller, vermiform borings are similar in size to *Girvanella* tubules suggesting that they may have been produced by the boring and perforating action of the algae.

The carbonate mud matrix consists of micrite, microspar and rare pseudospar (terminology after Folk 1965), the last two of which form subhedral to anhedral crystals with few inclusions. Interstices are filled with two major types of fine- to medium-crystalline calcite cement. Type 1 is intergranular and intragranular cement which consists of a first stage of nonferroan, radial-prismatic calcite and a second stage of nonferroan, equant calcite respectively showing mottled dull-moderate and dull luminescence. The second- stage calcite is best developed as inclusion-poor, syntaxial overgrowths on echinoderms. Type 2 is equant, ferroan and nonferroan calcite filling vugs up to one centimetre in diameter and fractures which crosscut all previously described features. Under cathodoluminescence, this calcite shows concentric zonation which, from rim to centre of the vug or fracture, changes from dull- to non- to bright- to dull-luminescence. Minor intergranular cements of isopachous, radial-fibrous calcite and inclusion-rich, syntaxial overgrowths on echinoderms predate not only the two major types of cement but also some carbonate mud matrix.

Wavy, subhorizontal dissolution seams occur in argillaceous laminae in mottled limestones. Stylolites are rare but, when present, are well developed and up to one centimetre in height.

DOLOMITE PETROGRAPHY

Dolomite occurs throughout all four limestone lithofacies but is highly fabric-specific. It occurs preferentially in micritic components which include (Fig. 6; 7): 1) the intergranular matrix in stromatoporoid-coral boundstones and intraclast floatstones-rudstones; 2) the internodular matrix in nodular limestones; 3) micritized portions of allochems, and 4) borings and burrow structures. The abundance of dolomite in each of these components is highly variable, ranging from less than 5 percent to 100 percent by volume (visual estimates). Intraclasts and unmicritized bioclasts remain as calcite but dolomite rhombs locally impinge on particle margins. Mottled limestone beds interpreted to be hardgrounds are also never dolomitized.

In mottled and nodular limestones, the transition between dolomitized and undolomitized areas is typically gradational; no other related changes in composition or fabric have been observed. In places, elongate allochems such as brachiopods and tabular stromatoporoids separate dolomitized from undolomitized areas. Some limestones, however, have dolomitized areas which are argillaceous and display distorted allochems and grain-contact sutures.

Two stages of dolomite are recognized in most Elm Point limestones. The first, more extensive stage consists of disseminated, fine-crystalline rhombs which are characteristically nonferroan and inclusion-rich, and exhibit a mottled dull-red luminescence (Fig. 7b). Dolomite rhombs locally form a mosaic with gently sutured crystal boundaries. They are truncated by dissolution seams as well as by vugs and fractures which are open or filled by type 2 calcite cement. Adjacent to some vugs and fractures, a second stage of inclusion-free, nonferroan dolomite is well developed as overgrowths on first-stage dolomite. It has zoned, moderate orange-red luminescence.

Stage 1 dolomite rhombs and adjacent intercrystalline matrix contain micromillimetre-sized, irregularly-shaped vugs indicating partial dissolution of the dolomite. Some vugs are completely or partially occluded by ferroan, dull-luminescent calcite. Stage 2 dolomite rims on the larger dolomite rhombs are essentialy unaltered except where adjacent to large calcite-filled vugs and fractures, when they are rounded and presumably corroded. Rhomb-shaped calcite is common in mottled limestone, particularly in the Steep Rock area in the northern part of the study area, indicating nearly complete replacement of dolomite by calcite (*i.e.* dedolomitization).

TIMING OF DOLOMITIZATION

Based on spatial and temporal relationships between calcite and dolomite in dolomitized limestones of the Elm Point Formation, a tentative paragenetic sequence has been established (Table 1). Mineralogical transformations of original aragonite and high-magnesium calcite, carbonate mud matrix lithification and mechanical compaction have not been included because suitable data are lacking.

Table 1: Summary of the preliminary paragenetic sequence for Elm Point limestones.

	SYNSEDIMENTARY	EARLY DIAGENETIC	LATE DIAGENETIC
* DEVELOPMENT OF MICRITE ENVELOPES	———		
* PRECIPITATION OF FIBROUS CALCITE CEMENT & SYNTAXIAL OVERGROWTHS	———		
* HARDGROUND FORMATION, INTRACLASTS	———		
* PRECIPITATION OF PRISMATIC & EQUANT CALCITE CEMENTS (TYPE 1)		———	
* DOLOMITIZATION (STAGE 1) OF MICRITIC COMPONENTS		———	
* FRACTURE & VUG FORMATION			———
* DOLOMITIZATION (STAGE 2)			———
* PRECIPITATION OF EQUANT CALCITE CEMENT (TYPE 2)			———
* DEDOLOMITIZATION			———
* PRESSURE DISSOLUTION			———

The first phase of diagenesis involved microbial micritization of allochems, precipitation of fibrous calcite cement and formation of hardgrounds and micritic intraclasts. It was probably synsedimentary, with alteration occurring in shallow-marine waters contemporaneously with or shortly after sediment deposition (*cf.* James and Choquette 1983). The second phase included fabric-specific processes of inter- and intraparticle (type 1) calcite cementation, followed by dolomitization (stage 1) of micritic components. These processes appear to have occurred prior to extensive compaction, *i.e.* during early diagenesis. The final phase of diagenesis included fracturing and dissolution which crosscut all earlier features. The fractures and vugs were filled by dolomite (stage 2) which precipitated as overgrowths on older dolomite (stage 1) and by a later calcite cement (type 2). Subsequent pressure dissolution and dedolomitization seem to be the last processes for which evidence exists in the Elm Point Formation.

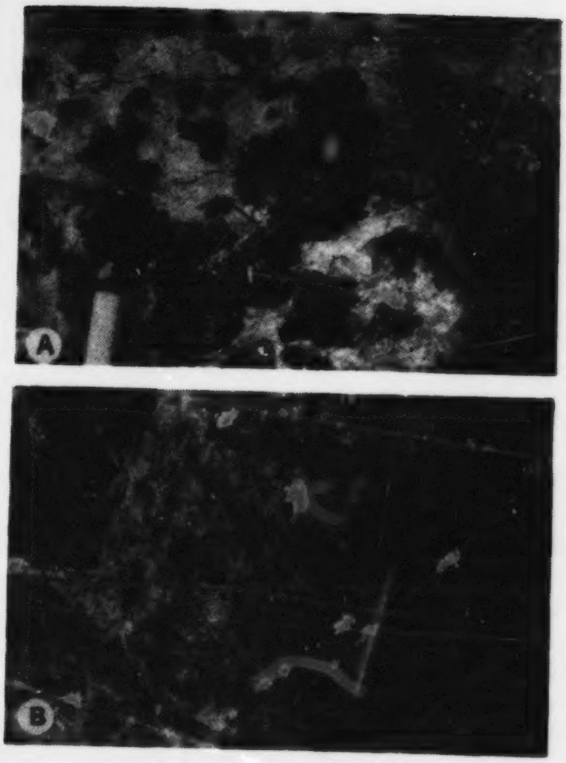

Figure 4: Outcrop photographs. A) Nodular, skeletal wackestone-packstone with limestone nodules (medium grey) in a tan-weathered dolomitic matrix (pale grey). B) Bedding plane view of mottled, skeletal wackestone-packstone showing tan-weathered, dolomitized burrows (medium grey, arrows).

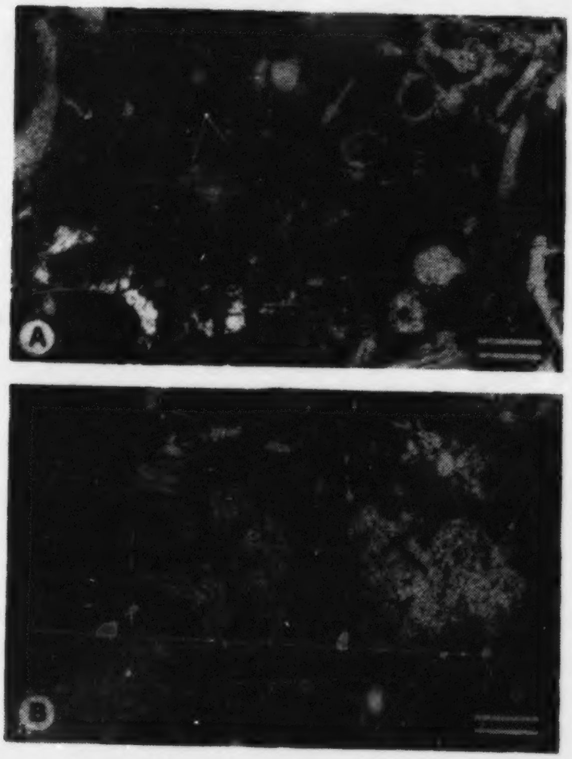

Figure 5: Thin-section photomicrographs. A) Mottled, skeletal packstone composed of a diverse assemblage of bioclasts; *Girvanella* intraclast is arrowed. Scale bar is 200 µm. B) Mottled, skeletal packstone showing micrite-filled borings in bioclasts (arrows). Scale bar is 200 µm.

Further refinement of this paragenetic sequence, particularly the character of the calcite cements and dedolomite, and interpretation of the mechanism(s) of dolomitization await further study.

DISCUSSION: INTRINSIC CONTROLS ON DOLOMITIZATION

Bioturbation and preferential dolomitization of burrows account for the texture of various nodular and mottled limestones in Paleozoic strata (*e.g.*, Kendall 1977; Demicco 1983; Moller and Kvingan 1988). These processes likewise occurred in the Elm Point Formation but other factors also controlled dolomitization. These factors are fabric-specific and include: 1) original sediment texture; 2) macroscopic borings; 3) microbial micritization, and 4) early limestone lithification. They locally affected sediment permeability and porosity which, in turn, controlled the extent of subsequent dolomitization.

The influence of original sediment texture on dolomitization is best demonstrated in skeletal wackestones which have elongate allochems such as brachiopods. In places, these allochems appear to act as local permeability barriers and separate dolomitized from undolomitized areas; although mineralogically distinct, these areas are texturally similar.

In addition to burrows, micrite-filled borings and micrite envelopes may be partially or completely dolomitized. This suggests that boring activity and microbial micritization enhanced sediment permeability and porosity which promoted localized dolomitization. In contrast, micritic intraclasts and layers probably representing submarine hardgrounds have never been observed to be dolomitic. They were apparently synsedimentary in origin and suggest that early lithification effectively reduced sediment permeability and porosity which, in turn, inhibited dolomitization. Assessment of the extent and influence of early lithification in other micritic constituents that are undolomitized, such as limestone nodules in nodular limestone, is inconclusive.

Some studies of Cenozoic carbonates have shown that precursor mineralogy and particle or crystal size are important controls on dolomitization. Commonly, high-magnesium calcite and aragonite components, as well as lime muds, are preferentially dolomitized (*e.g.*, Murray and Lucia, 1967; Buchbinder and Friedman, 1970; Land, 1973; Sibley, 1982). Elm Point limestones lack evidence for this kind of mineralogical or size control on dolomitization. Hence components such as stromatoporoids (*cf.* Rush and Chafetz 1991) and, probably, echinoderms, which were originally composed of high-magnesium calcite, are preserved as calcite. Also, lime mud in the Elm Point Formation is not consistently dolomitized.

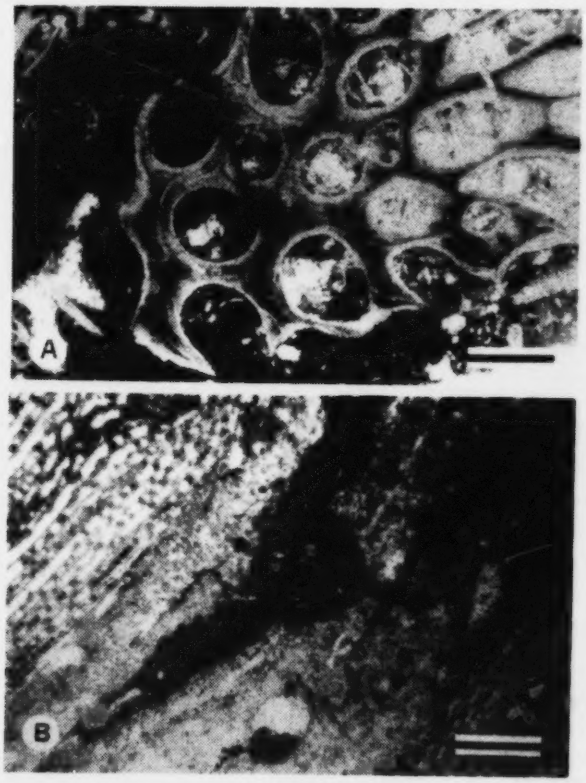

Figure 6: Thin-section photomicrographs. A) Well-preserved bryozoan in dolomitized matrix (dark grey); vugs (lower left corner) and some intraparticle voids in the bryozoan are filled by equant calcite cement (white). Scale bar is 1 mm. B) Preferentially dolomitized boring (arrow) in tabular stromatoporoid. Scale bar is 2 mm.

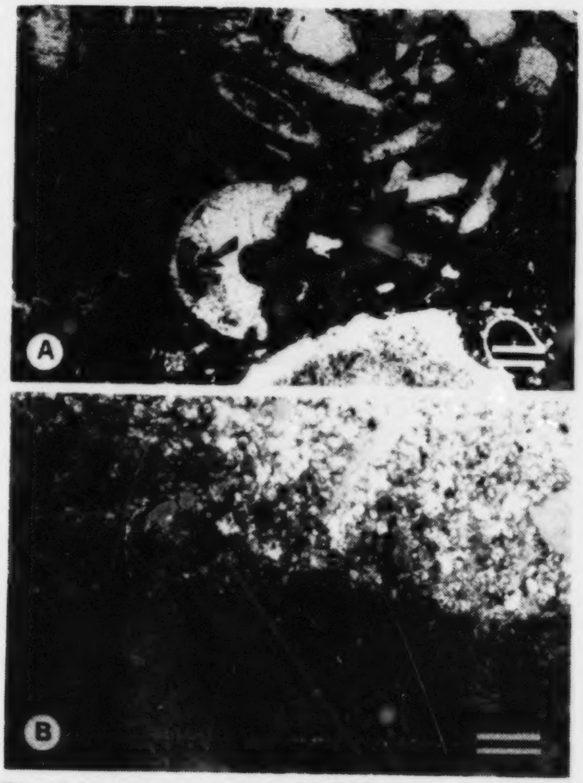

Figure 7: Thin-section photomicrographs. A) Calcite bioclasts in dolomitized matrix; crinoid ossicle contains preferentially dolomitized boring (arrow). Scale bar is 400 µm. B) Intraclast floatstone showing portion of micritic intraclast (medum grey, lower half) and dolomitized interparticle matrix (pale grey, upper half); skeletal fragments in intraclasts are truncated by intraclast margins (arrows); dolomite rhombs (white, upper half) are scattered throughout the matrix. Scale bar is 200 µm.

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GROWTH PATTERNS OF THE MIDDLE DEVONIAN WINNIPEGOSIS FORMATION, BLUFF REEF DAWSON BAY AREA, MANITOBA

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ABSTRACT

The outcrops of the Bluff Reef are the most accessible Middle Devonian reef exposures in western Manitoba. The Bluff is a near shore island on the northwest side of Dawson Bay. Unlike other island exposures of the Winnipegosis that are accessible only by boat or float-equipped aircraft, the Bluff is connected to the mainland by a tombolo of Holocene and Pleistocene deposits. The reef outcrops are on the north and northeast margin of the island, but recent corehole drilling at the south end indicates that the shape of the island closely imitates the outline of the Winnipegosis reef. However, the southern half is overlain by 26 to 38 m of Dawson Bay Formation, Second Red Beds and a collapse breccia. Thus, at the north end the reef is about 70 m thick, but in the south half it varies from 43 to 51 m. The reef has a northeast axial trend and a surface length of about 700 m along that axis.

A seven-corehole drilling program indicates that the Bluff Reef rests conformably on the same carbonate rock unit that underlies all Winnipegosis reefs, commonly called the platform. The platform consists of mottled to nodular dolomitized lime mudstone with scattered crinoid columnals and fragmented brachiopod valves. The platform in this area has a variable thickness ranging from about 8 m at both ends of the Bluff to 14 m beneath the central portion.

At the south end of the reef, growth was initiated through a proto-reef community, similar to that found forming the foundations of the Swan Hills reefs of Alberta. This community was dominated by small hemispherical stromatoporoids and thamnoporid corals. They occur in a dolomitized micritic rock with a large proportion of kerogenous material. At the northern end of the reef, the inaugural community appears to have been dominated by crinoids. With the exception of a second proto-reef community a few metres above the founding proto-reef, the remainder of the vertical sequence at the Bluff is dominated by dolomitized micrite containing mainly metozoan skeletal remains as well as peloids and calcareous algal debris. Coreholes and outcrops show that the upper 15 to 30 m at the northern end consist of tabular hemispherical and bulbous forms of stromatoporoids, as well as thamnoporid, fasciculate and chaetitid corals and stringocephalid brachiopods.

Three conclusions can be drawn about Winnipegosis reefs from this study of the Bluff reef: 1) Winnipegosis reef morphology can be extremely changeable; 2) although most of the lithologic variations occur in a vertical sense, there are some subtle and not so subtle lateral changes; and 3) growth of the reef appears to have been initiated about some irregular form on the Middle Devonian sea floor.

INTRODUCTION

Geologists have been frequent visitors to the Bluff through the years. The presence of Devonian rocks on Lakes Manitoba and Winnipegosis was established as early as 1859. However, it was left to J.B. Tyrrell (1892) to establish a stratigraphic succession. He subdivided the sequence into a Lower Devonian red bed, a Middle Devonian Winnipegosan Formation and an Upper Devonian Manitoban Formation. Baillie (1951) was the first to recognize that the local positive topographic relief along the shores of both Lakes Manitoba and Winnipegosis, as well as many of the islands in Dawson Bay, are genetically related to reefs of the Winnipegosis Formation.

Since the late 1940s when oil exploration in western Canada intensified, the Bluff has been thought of as a good, accessible example of a Winnipegosis reef. Its biota and lithologies are considered to be typical reefs in the outcrop belt and the components observed at the surface were thought to continue to depth. Since most of the subsurface Winnipegosis reefs did not appear to contain the same biota and lithologies over their total thickness, many subsurface geologists believed that there were some major differences between the outcropping and subsurface Middle Devonian reefs of western Canada.

A corehole drilling program, initiated by the Manitoba Geological Services Branch in 1969 and continuing to the present, has led to a much improved understanding of outcropping and near surface Winnipegosis reefs. The raw data from this drilling program were published in annual reports of Field Activities released by the Manitoba Department of Energy and Mines. Interpretations of the results of the program have been published by Norris *et al.* (1982) and McCabe (1986, 1987).

Due to its accessibility, the Bluff has become the focal point for a detailed investigation. The study has been carried out by us with support from Manitoba Energy and Mines, a consortium of oil companies (Amoco Canada, Chevron Canada Resources, North Canadian Oil, PetroCanada and Saskoil) and, most recently, the Potash Producers of Saskatchewan. Bluff data plus information from two other reef locations are the basis for a master's thesis by the junior author. To avoid pre-empting the thesis, this paper will focus mainly on observations, leaving much of the interpretation to the thesis.

An initial five coreholes were drilled at the north end of the Bluff in the summer of 1988. Two of these penetrated to the Ashern Formation, the others penetrated some 45 to 54 m of reef and were terminated because of unconsolidated quartz sand at that depth which made coring impossible. Two additional coreholes were drilled in the spring of 1990, one at what was thought to be a mid-reef location and the other at its south end. In addition, all outcrop localities at the Bluff were examined in detail and were referenced to the coreholes by a horizontal and vertical survey employing a plane table, alidade and stadia rod.

Lying some 340 km northeast of Regina, Saskatchewan, and a similar distance northwest of Winnipeg, Manitoba (Fig. 1), the Bluff is located on the western shore of Dawson Bay, a large embayment at the northern end of Lake Winnipegosis. The nearest large settlement is Swan River, Manitoba, about 100 km south on Manitoba Highway 10. Using a survey monument located on the northeast shore of the Bluff, the site can be referenced to the western Canadian land survey grid and is determined to be in Lsds. 1, 2, 7, 8-35-45-25W1 and 5-36-45-25W1, respectively.

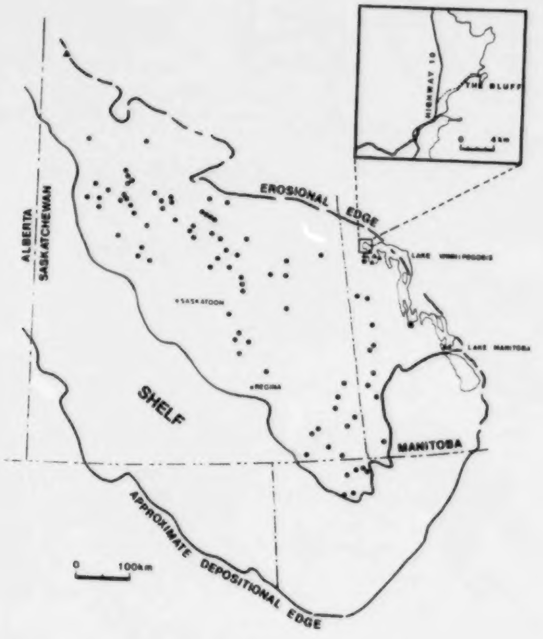

Figure 1: Location map. The map shows the simplified paleogeography of the Elk Point Basin the position of the Bluff Reef within it. The inset map shows the geographic location of the Bluff with respect to Manitoba Highway 10 north of Swan River, Manitoba.

The Bluff is a near shore island, connected to the mainland by a tombolo of Quaternary alluvium and is accessible from Manitoba Highway 10 along a graded road and a well-travelled bush trail (Fig. 1). The shoreline of the island traces out an oblong shaped landform with a northeast axial length of about 700 m (Fig. 2). Major topographic relief on the island is along the northern and northeastern shore where there are 30 and 40 m lengths of continuous outcrop in cliffs 6 to 8 m high. Difference in relief from the northern to southern end never exceeds 1.5 m. The Bluff is extensively tree-covered, but much of the undergrowth has been grazed-off by livestock. Thus, in addition to the cliffs, there are numerous low relief surface exposures.

When the initial corehole program was developed, it was thought that the northern end of the Bluff was the only part underlain by a Winnipegosis reef and the location was interpreted as a small diameter pinnacle reef. The identification of boulders in the surficial deposits at the southern end as having been sourced from the Dawson Bay Formation led to drilling of the sixth and seventh coreholes. The assumption that the Dawson Bay Formation was held up by a part of the Bluff Reef with lower relief than that of the northern end was verified by both coreholes.

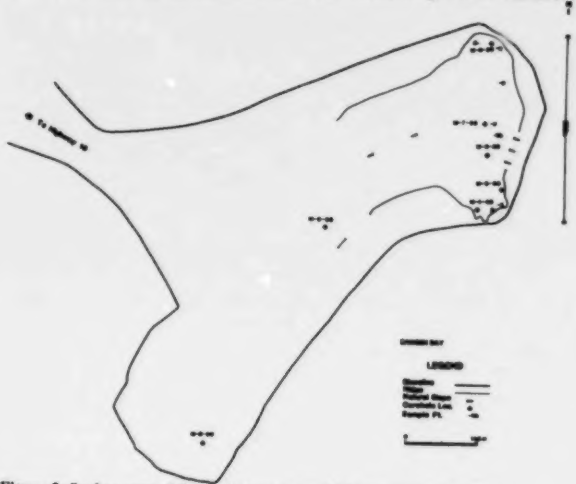

Figure 2: Surface map of the Bluff showing corehole and outcrop locations.

REGIONAL GEOLOGICAL SETTING

The Bluff area lies on the gently dipping southwestern limb of the Severn Arch; the gradient of the Precambrian surface in the area is about 2.3 m/km. The area immediately overlies a gravity anomaly in the Precambrian basement that appears to delineate the western edge of the Superior Province and eastern margin of the Trans-Hudson orogen.

Based on the structure contour maps of Norris *et al.*, (1982) we estimated that the Bluff area is underlain by some 220 m of Lower Paleozoic rocks. These consist of basal clastic strata belonging to the Winnipeg Formation and a succession of carbonates, mainly dolomite, that make up the Red River, Stony Mountain, and Stonewall Formations and the Interlake Group. The latter is immediately overlain by the Ashern Formation, the oldest strata penetrated by the Bluff coreholes.

The rocks of the Ashern Formation lie with slight angular unconformity on the Interlake Group and are the oldest Devonian strata in Manitoba. They are thought to be Eifelian in age (Norris *et al.*, 1982). Regionally, the Ashern varies from about 3 to 12 m in thickness. None of the Bluff coreholes penetrated the full Ashern, but several coreholes about 11 km south of the area include sections of about 8 m. Lithologically, it is a remarkably persistent unit, composed of reddish brown to greyish orange and greyish green, argillaceous dolomite. The greenish coloration commonly occurs at the top of the formation near its contact with the overlying Winnipegosis. Angular clasts of yellowish grey, microcrystalline, dolomite, typical of the Interlake, commonly occur in the basal beds. The basal contact is fairly sharp because of the distinct difference in lithology between the Interlake and Ashern. By contrast, the upper contact is less well defined due to an intermixing of the argillaceous sediments of the Ashern with the overlying non-argillaceous carbonates of the lower Winnipegosis.

Throughout Manitoba, the Ashern is overlain by the Elm Point Formation or the lower Winnipegosis (Fig. 3). The former name is restricted to rocks composed of high-calcium limestone and the latter term is applied to more dolomitized strata. Both are mottled lime mudstone to wackestone with a typical normal marine biota of scattered colonial corals, bulbous stromatoporoids, brachiopods, crinoids, gastropods and conodonts. According to Norris *et al.* (1982), the intensity of dolomitization in these rocks is greatest beneath the Winnipegosis reefs. The thickness of the Elm Point/lower Winnipegosis strata varies from 11 to 20 m. The nature of its contact with the upper Winnipegosis also varies depending on whether or not it is overlain by a reef complex or inter-reef rocks. Beneath the reefs the contact is placed at a change from mottled, yellowish grey or yellowish brown carbonate to light yellowish grey, crinoidal wackestone. In the inter-reef areas, the overlying rocks are generally dark yellowish grey, thinly bedded to laminated, with hairline-thin kerogenous partings emphasizing the bedding and interlaminar interfaces.

McCabe (1986, 1987) showed that the Winnipegosis reefs in the outcrop belt of western Manitoba are an assortment of sizes and shapes (Fig. 4). They range from the low relief mounds (40 to 50 m) of the Lake Manitoba Narrows area to the higher relief (65 to 88 m) flat-top and pinnacle reef complexes of the Lake Winnipegosis area. The Bluff belongs to the latter of these. McCabe (1986) interpreted the size and shape of the reefs to be related to their paleogeographic location in the shelf to basin setting of the Middle Devonian Elk Point Basin (Fig. 5). The low relief mounds probably were formed high on the slope immediately beneath the shelf break. By contrast, the flat-top and pinnacle reef complexes are thought to have grown in a deeper basin setting.

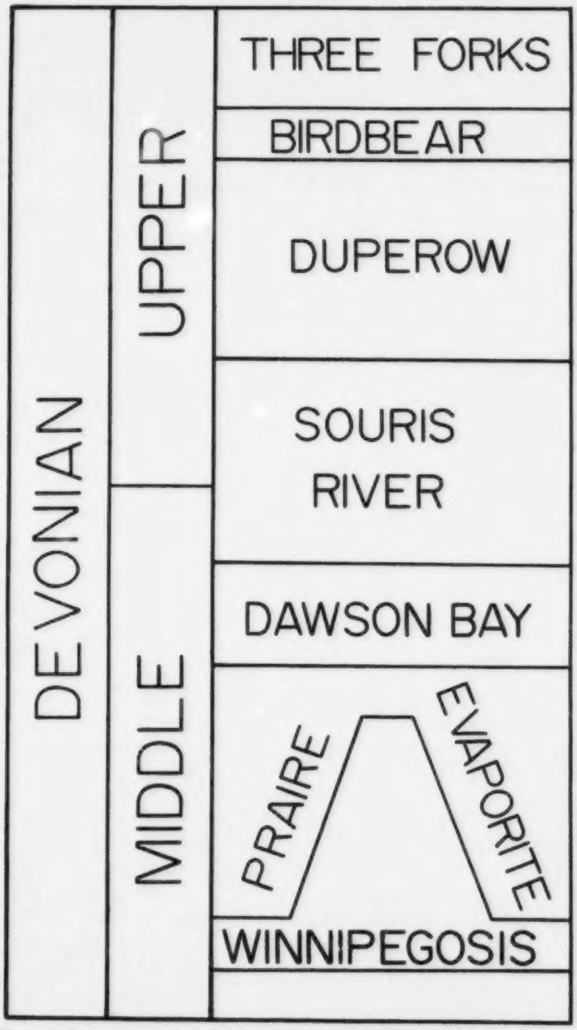

Figure 3: Nomenclature chart of Devonian rocks in Manitoba and Saskatchewan.

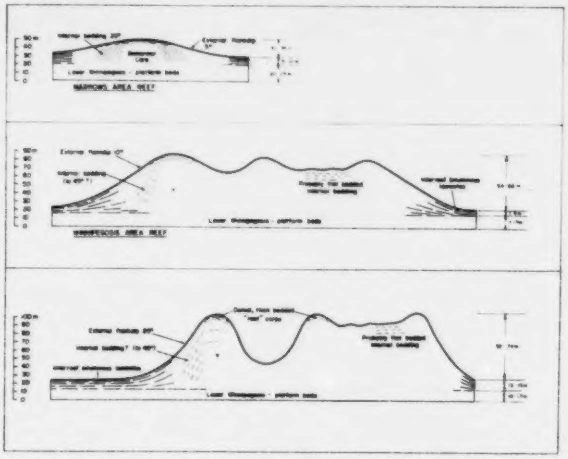

Figure 4: Schematic illustrating the varying geometry of Winnipegosis reefs in the Manitoba outcrop belt sections (McCabe, 1986).

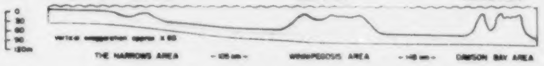

Figure 5: Schematic cross-section of the shelf to basin setting of in the Middle Devonian Elk Point Basin of Manitoba (McCabe, 1986).

Southwest of the study area the Winnipegosis reefs are enclosed and overlain by halite and anhydrite belonging to the Prairie Evaporite (Fig. 3). However, in most of the region immediately west of Lakes Manitoba and Winnipegosis, the Prairie Evaporite has been removed by solution (McCabe, 1967) and the strata overlying it, Dawson Bay and Souris River, were lowered onto the Winnipegosis, draping the reefs as anticlinal structures and filling the inter-reef areas. In some places, such as the Bluff, Pleistocene glacial advances stripped the younger Devonian strata from the highest portion of the reef leaving younger strata as inliers between the reefs. Salt solution and younger stratal draping are important clues to Winnipegosis reef size and shape in the outcrop belt because the drape structures are commonly expressed as positive rises (McCabe, 1986, 1987) and their surface expression can be used to help interpret reef morphology.

VERTICAL AND LATERAL GROWTH PATTERNS OF THE BLUFF REEF

The vertical and lateral growth patterns of the Bluff Reef are based on apparent changes in the reef biota. The idea of ecologic successions in Phanerozoic reefs has been presented by an assortment of researchers and the topic was recently reviewed by Copper (1988). He points out that for many Phanerozoic reefs, a four-fold ecologic succession can be recognized, *i.e.*, stabilization, colonization, diversification and domination, but for others three stages appear to be more applicable. The Bluff Reef fits into the latter category and we propose a mixture of previously applied phase names. Thus, the initial stage is identified as stabilization, stage two as colonization and stage three as climax.

Figure 6 is our interpretation of the morphology of the Bluff Reef. It is evident that the southern portion of the reef did not attain its fullest relief and vertical growth was, in fact, terminated at some time during the second stage. By contrast, the north portion of the reef continued to grow to the climax stage. Figure 7 has been used as a base for our ecologic succession maps.

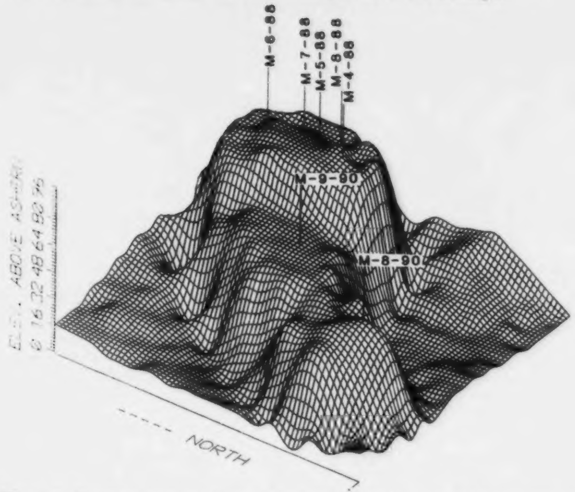

Figure 6: Computer-generated three-dimensional model of the Bluff Reef. Vertical exaggeration is 5.4x.

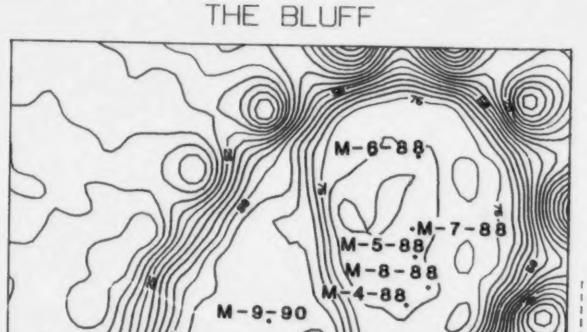

Figure 7: Map demonstrating the approximate shape of the Bluff Reef and the locations of the coreholes. The contours are in metres above the top of the Ashern Formation and conform to the upper surface of the reef.

STABILIZATION STAGE

Figure 8a illustrates the stabilization stage and is marked by a biofacies variation from south to north. The illustration shows a dominance of thamnoporid corals and small bulbous stromatoporoids in the southern half and a preponderance of crinoids in the north. However this zonation may be more apparent than real, as Copper (1988) suggested that biotic zonation in pioneer communities is uncommon and, also, there is rather poor deep borehole control at the north end of the reef, where none of the holes (M-5-88, M-6-88 and M-7-88) penetrates to the base of the reef. Brachiopod valve fragments, crinoid columnals and scattered fronds of bryozoans make up the other allochemic debris of the southern half. All allochems are in a micritic matrix and seldom constitute more than 30 percent of the total composition of the rock.

According to Copper's (1988) model of the pioneer stage of a Siluro-Devonian reef, the allochems identified in the base of the Bluff Reef are typical of the organisms found in the initial growth stage of reefs of that time period. He described them as "opportunists", *i.e.*, quite capable of establishing themselves on newly developed substrates.

The factors that influence the pioneer communities of Winnipegosis reefs to establish themselves at a particular site have not been clearly determined. At the Bluff, the reef appears to be located on a lower Winnipegosis platform of variable thickness. Coreholes M-4-88 and M-9-90 penetrated about 14 m of platform. According to Norris *et al.* (1982) the scope of regional thicknesses for the platform is 11 to 17 m. A transect of coreholes at the mouth of the Steeprock River, 23 km south of the Bluff, has a thickness of approximately 16 m for the lower Winnipegosis (McCabe, 1987). Therefore 14 m of lower Winnipegosis in the Bluff area is similar to regional thickness. However, coreholes M-8-88 and M-8-90 penetrated a total thickness of only 8 m of platform rock, suggesting thereby draping of some form of positive relief at these two sites. This positive relief may have prevailed during the inauguration of reef growth. One may speculate on the origin of these positive features. One possibility is paleokarstic hills on the Interlake rocks. An opposing argument would be that the 14 m of platform is greater than regional thickness for the area, and the stabilization community was established on the flanks of a paleobathymetric rise caused by the thicker platform. Rosenthal (1987) identified a similar initial setting for a Winnipegosis reef in the vicinity of the town of Winnipegosis at the south end of Lake Winnipegosis.

If positive elements were at the sites of M-8-90 and M-8-88, then the paleobathymetric rise at M-8-90 was the locale of growth of thamnoporid corals and small bulbous stromatoporoids. By contrast, the feature penetrated by M-8-88 acted as the nucleus for a crinoid meadow. The lithologies at both sites are wackestones to wacke-packstones with closely spaced kerogenous stylolites (stylocumulates). The stylocumulates may be evidence for concentrated chemical solution at each of these sites, as might be expected over a positive element. Whichever scenario is valid, it appears that the stabilization stage of the Bluff Reef was initiated on one or more knolls on the Middle Devonian sea floor.

COLONIZATION STAGE

The colonization stage of the Bluff can be considered as a transitional growth phase, in which the diversity of biota markedly increases. Thus, Figure 8b shows brachiopods, bryozoans, tabulate and fasciculate corals, crinoids and ostracods as the main fauna, and a flora that includes codiacean (*Lancicula* and *Litanaia*) and dasycladacean algae, as well as microbialites (used in the sense of Burne and Moore, 1987). The brachiopod remains are mainly fragmented but appear to be dominantly stringocephalids. Recognizable coral debris in thin section includes both thamnoporid and disphyllid forms, but the former are more abundant. The dasyclads are represented by calcispheres, and microbialites are commonly coatings of *Sphaerocodium* or *Stenophycus* on bryozoan and coral debris.

Peloids are the main nonskeletal material and in places they constitute peloidal grainstones to packstones with scattered biological allochems intermixed. In places, thin sections of the peloidal grainstones and packstones show them to have a layered appearance. Elsewhere, there are mixtures of nonskeletal and skeletal allochems that compose wackestones to wacke-packstones. Where present, the matrix material is dolomicrite.

The tabulate and fasciculate corals along with the calcareous algae probably acted as sediment baffles, and the high proportion of lime mud that accumulated because of this baffling process made the site more conducive to tabulate coral growth (Jull, 1977). On the other hand, the fasciculate corals were probably restricted to the margins of the buildup and to more stable substrates such as those supplied by grainstones, packstones or subaqueously cemented lime muds (Jull, 1975).

CLIMAX STAGE

Copper's (1988) idealization of a Siluro-Devonian climax reef indicates domination by organisms that develop large colonial skeletons, as well as specialist bivalves and diversified crinoids. At the Bluff the climax biota includes hemispherical, encrusting and tabular stromatoporoids; fasciculate, tabulate and chaetitid corals; stringocephalids, crinoids, calcareous algae, microbialites and cryptobionts (Fig. 8c).

Peloids and intraclasts compose the nonskeletal fraction of the reef. All allochems, both skeletal and nonskeletal are in a matrix of dolomicrite that in varying proportions result in combinations of wackestones, wacke-packstones and packstones.

In the southeast portion of the climax reef, the stromatoporoids are commonly in growth position and, in places, appear to have been established on substrates formed by horizontal clusters of stringocephalids. Elsewhere, the fasciculate corals use the same form of substrate. The distribution of both faunas suggest a prevailing wave surge from the east-northeast. The cryptobionts are most easily recognized in *Renalcis* or *Renalcis*-like forms and *Epiphyton*; on the other hand the microbialites are *Sphaerocodium* and *Stenophycus* coatings of coral debris and lithoskels, *i.e.*, intraclasts of microbialites. The microbialite lithoskels appear to be more common at the extreme north end of the reef where the strata apparently dip northward. Thus, they may have been eroded from the interior of the reef during its growth and deposited in the flanking beds.

The presence of microbialites in both the colonization and climax stages of growth may have some implications on water depth at the time of vertical growth. Jones and Hunter (1991) report microbialites as part of a community replacement sequence in the Pleistocene carbonates of Grand Cayman Island and they interpret them as indicators of regressive or shallowing conditions.

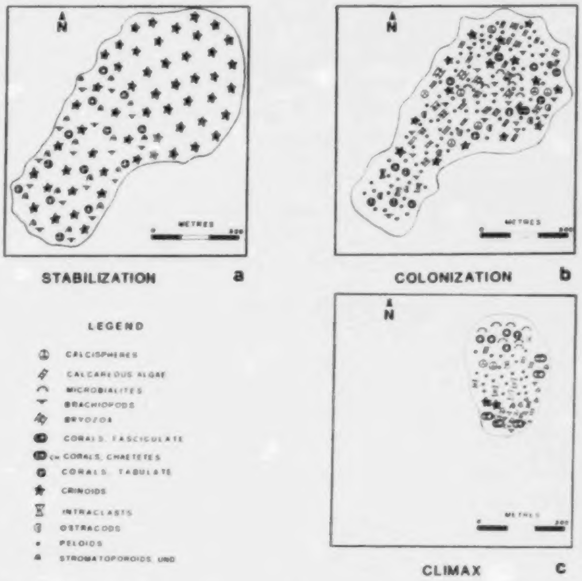

Figure 8a: The stabilization stage of growth, represented by the lithologies in a 10 to 15 m interval immediately overlying the lower Winnipegosis platform rocks.

Figure 8b: The colonization growth stage, represented by lithologic data taken from strata about 10 m thick, 30 m above the Ashern Formation.

Figure 8c: The climax growth stage, represented by lithologic data from a 10 m interval 70 m above the Ashern Formation.

COMPARISON OF OUTCROPPING AND SUBSURFACE REEFS

Fay and Copper (1982) and Copper and Fay (1988) demonstrated that reefs within a specific depositional basin may not necessarily have replicate ecologic successions, although growth patterns may follow a general trend. They found in particular that the climax communities differed with paleogeography. Thus, to expect mirror images in outcropping and subsurface Winnipegosis reefs may be a simplification of the facts. Observations by Rosenthal (1987) also emphasize this point; he noted that only one reef, Rock Island, in his study area contained typical stromatoporoid-coral framestones and bindstones in the climax phase. Notwithstanding the previous comment and Rosenthal's observations, it is important to remember also that any apparent differences between outcropping and subsurface Winnipegosis reefs may be related to the site on the subsurface reef from which a core was taken. For example, Perrin and Precht (1985) did not recognize a climax community at the top of the cored North Dakota reefs. They interpreted the rocks to represent an intra-reef lagoonal facies. It is quite likely that boreholes drilled elsewhere on those reefs would have penetrated a climax community. Regardless of these disparities, some broad comparisons and contrasts can be drawn, and five recent papers, Martindale and MacDonald (1990), Martindale and Orr (1988), Eherts and Kissling (1987), Gendzwill and Wilson (1987), and Perrin and Precht (1985), are useful in making them.

We indicated that it is not clear to us what controls the site of growth of the pioneer community for a Winnipegosis reef. However, the biota of the stabilization phase of both outcropping and subsurface reefs appear to be similar. We have observed an excellent example of the crinoidal pioneer community in the core of the Amerada Tenneco Langbank 15-28A well (Lsd. 15-28-12-2W2). It occurs over a 10 m interval from 1696 to 1706 m, and the lower 2 m is remarkably similar to M-8-88 at the Bluff, in that crinoid columnals are concentrated in stylocumulates at both localities. Perrin and Precht (1985) also identified crinoids as a major contributor to the stabilization stage of reefs in the subsurface of North Dakota, but there they appear to be associated with abundant brachiopods. This is well illustrated by their Figure 17A, showing a typical wispy nodular platform carbonate in sharp contact with a crinoidal-brachiopodal packstone. The latter represents the basal community of the overlying reef. In addition, Kissling and Eherts (1987) and Martindale and Orr (1988) identified crinoids in the basal strata of other subsurface reefs. The latter have also noted that stromatoporoids are part of the stabilization community of at least one subsurface reef.

Most authors recognize a similar colonization phase. Eherts and Kissling (1987) and Martindale and MacDonald (1990) refer to it as a peloidal/codiacean algal facies, an appropriate description based on the main components of this growth stage. However, Perrin and Precht (1985) recognized a colonization phase of *Thamnopora*, *Favosites*, *Hexagonaria* and tabular stromatoporoids which is followed in vertical sequence by their interpreted lagoonal facies. Gendzwill and Wilson (1987) developed a composite Winnipegosis reef model that included the peloidal rocks of the transition phase and Perrin and Precht's (1985) lagoonal facies. Accordingly, from their observations they found no climax community.

Eherts and Kissling (1987) distinguished a climax community remarkably similar to that present at the Bluff. However, the Tableland Reef climax community (Martindale and MacDonald, 1990) is dominated by tabulate corals, red algae and microbialites of *Sphaeracodium*. Chaetitid corals also occur in one of the wells of that reef.

In summary, the main similarity between growth patterns of Winnipegosis reefs, whether in outcrop or subsurface, is in the stabilization and colonization phases. The climax phase may differ, but what is interpreted from core may be strongly influenced by the borehole location on the reef. The major difference between the Bluff Reef and subsurface reefs such as Tableland, where there is sufficient control to determine their geometry, is in their areal extent. The Bluff has an interpreted basal length of 916 m and a maximum interpreted basal width of 460 m. By contrast, Tableland is about four and a half times longer and three times as wide. In addition, climax reef development was restricted to the northern end of the Bluff; accretion of sediment on the south part of the reef appears to have ceased some time in the colonization phase. This contrasts with Tableland where no differential growth is evident.

CONCLUSION

McCabe (1986, 1987) has shown that there is an assortment of morphologies for Winnipegosis reefs in the outcrop belt of western Manitoba. The Bluff Reef is but another variant and it is likely that these morphologies are replicated in the subsurface of the Williston Basin. The tiered appearance of the Bluff may not be unique, and this outcropping reef may be useful as an exploration model for subsurface reefs with similar shapes. At least one subsurface reef in southeast Saskatchewan may be similar to the Bluff. At the Shell Oxbow 13-24-2-3 well (Lsd. 13-24-2-3W2) a section of Winnipegosis reef was cored which consisted of 38 m of mosaic, nodular and layered anhydrite resting on dolomitized wackestones to wacke-packstones of codiacean algae and other poorly preserved skeletal remains. This sequence suggests a colonization phase overlain by evaporites. A collapse breccia overlies the colonization phase of the southern half of the Bluff, suggesting that prior to dissolution of the Prairie Evaporite it too may have been overlain by anhydrite. Thus, the Bluff and the Oxbow well may represent similar depositional settings and, provided that the prevailing Middle Devonian wind direction was the same for both areas, it is possible that a thicker section capped by climax reef may lie a kilometre or so to the north of the existing well site. Consequently, evidence from this outcrop study could turn a site that apparently has poor hydrocarbon potential into a good prospect.

ACKNOWLEDGEMENTS

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WINNIPEGOSIS BUILDUPS OF THE HITCHCOCK AREA, SOUTHEASTERN SASKATCHEWAN - A CASE STUDY

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ABSTRACT

Since the discovery of significant quantities of oil in Winnipegosis (Givetian) buildups in southern Saskatchewan in 1986, intense drilling activity has proven that this play extends well beyond the area of the initial discovery at Tableland (Twp. 2, Rge. 9 W2M). This case study describes the sedimentology, diagenetic history, seismic configuration, geochemistry and productivity of four separate Winnipegosis buildups in the Hitchcock area (Twp. 3, Rges. 8 and 9W2M).

The Hitchcock buildups are founded on a regionally extensive Lower Winnipegosis platform. They are totally dolomitized, vary from 57 to 63 m in thickness and average 25 hectares (63 acres) in area. Buildups were located using 2-D seismic but their overall configuration and the well locations were determined by a subsequent 3-D survey.

The lower 40 to 50 m of each buildup consist of steeply dipping (>30° locally), porous peloid and codiacean algal packstones and grainstones. The overlying unit is dominated by chaetetids and is locally bound by "stromatolitic" algae and fenestrate bryozoans. An in-situ framework, 3 to 4 m thick, of red algae, colonial corals, fenestrate bryozoa and stromatolitic algae represents the final phase of organic reef growth. Buildups are capped by planar laminated or highly contorted calcretes. Laminar, nodular and brecciated dolomitic anhydrite (Whitkow Anhydrite) overlies the buildups and acts as a seal to entrapped hydrocarbons.

Buildups were initiated as shoals stabilized by early marine fibrous cements. Emergence and subaerial erosion terminated reef growth and produced a capping calcrete crust and a system of solution pores and cavities. The pores and cavities were subsequently infilled by mudstones of probable vadose origin. Dolomitization occurred by seepage reflux of marine waters during shallow burial.

Euxinic mudstones occur in basinal areas between buildups. They are probably coeval with the buildups and are organic rich and thermally mature. Their geochemical "signature" matches that of Winnipegosis oils in the area.

To January 1, 1991, after 7 to 15 months of production, four wells in the Hitchcock area flowed 40,120 m^3 (252,346 bbls) of oil.

INTRODUCTION

Dolomitized buildups of Middle Devonian (Givetian) age occur throughout the Elk Point Basin in Alberta, Saskatchewan, Manitoba and North Dakota. Despite a long history of exploration, no commercially significant quantities of hydrocarbons were produced from the Winnipegosis of Saskatchewan until 1986 when Home Oil and Canadian Hunter drilled the 8A-22-02-09W2M Tableland well just west of the town of Estevan in southeastern Saskatchewan (Fig. 1). Since that discovery, more than 60 wells have been drilled to the Winnipegosis in southeastern Saskatchewan, with a cumulative production to January 1, 1991 of 406,327 m^3 (2,555,715 bbls) of oil. The most intensive activity has been in the Hitchcock - Macoun area (Twps. 3 and 4, Rges. 8 and 9) north of Tableland. This case study will examine the local geological setting of four separate buildups in the Hitchcock area, their seismic expression, depositional and diagenetic history, the geochemical characteristics of the oil and source rocks, some of the more unusual completion practices employed and a number of the production problems which have been encountered in this area.

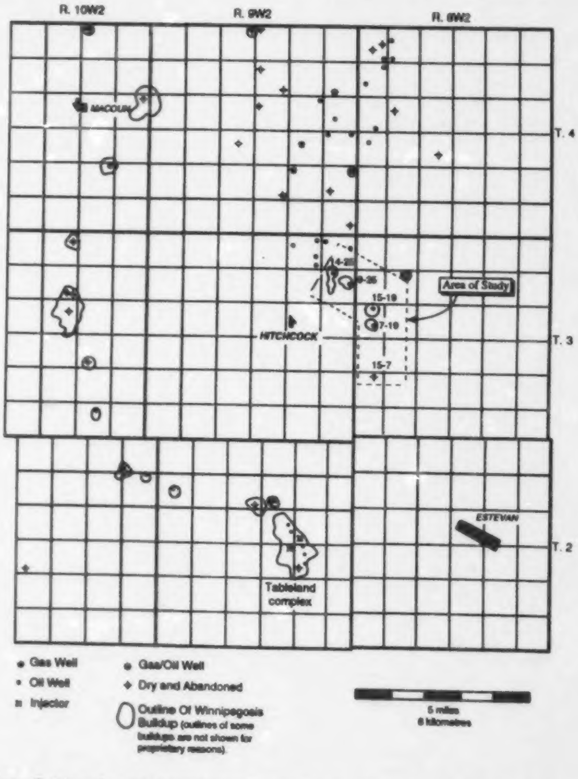


Figure 1: Map of southeastern Saskatchewan showing the location of Winnipegosis wells at Hitchcock and adjacent areas.

EXPLORATION HISTORY

Prior to 1968, numerous deep wells had established the presence of porous, dolomitized Middle Devonian buildups in the Elk Point Basin of Saskatchewan. Hydrocarbon discoveries in the equivalent Keg River Formation in northern Alberta in the 1960's led to renewed interest in the Winnipegosis and in 1968, a group of companies with substantial landholdings formed the Pheasant Consortium and drilled 49 Winnipegosis wells. Despite some oil shows, none of the Pheasant wells encountered commercial quantities of hydrocarbons.

In 1976, Dome Scurry Tableland (11-14-02-09W2M) was completed in the Winnipegosis. This well produced 5000 m^3 of oil before being abandoned in 1983 due to salt plugging in the well bore. In 1986, Home Oil Company Limited conducted a seismic survey over the Dome Scurry well and identified a structurally higher area to the northwest. The 8A-22 well flow-tested 223 m^3/day of 37° API oil from this feature. Since that time, the Tableland field has produced over 120,000 m^3 (750,000 bbls) of oil and over 20 productive wells have been drilled in the surrounding area.

It was against this background that Home Oil (via its subsidiary Scurry Rainbow Oil Ltd.) embarked on a program of seismic acquisition in the Hitchcock area in 1987. Early exploration in the Tableland area from 1985 to 1987 was conducted using conventional 2-D seismic surveys. Although Winnipegosis buildups can readily be identified using this technique, structural details and the areal extent have to be inferred between seismic lines. The inability of 2-D seismic to adequately delineate these features resulted in the drilling of a number of wells which either intersected the reef flank or missed the feature altogether. To overcome this problem, 3-D seismic data were used to detail the Hitchcock buildups once they had been located using conventional 2-D data. In addition, enhanced full 3-D migration techniques were found to substantially improve on conventional migration procedures.

REGIONAL SETTING AND STRATIGRAPHY

The Hitchcock area (Twp. 3, Rges. 8 and 9 W2M; Fig. 1) lies in the southeastern corner of Saskatchewan, approximately 26 km north of the border with the United States, and occupies a position in the southwestern corner of the Devonian Elk Point Basin. This basin extended from the Northwest Territories of Canada to the southern part of North Dakota and from the West Alberta Ridge in the west to beyond the erosional outcrop in the east (Fig. 2). The basin, which was barred from the open ocean to the north by the Presqu'ile Barrier, was fringed by a broad carbonate platform which enclosed a deeper, basinal area in which Winnipegosis buildups are located.

The basal Middle Devonian is represented by the Ashern Formation (Fig. 3) which varies from 10 to 16 m in thickness at Hitchcock and Tableland. It consists almost entirely of argillaceous, microcrystalline dolostone with no diagnostic body or trace fossils (Martindale and MacDonald, 1989). Lobdell (1984) suggested that the Ashern Formation was deposited in a restricted embayment in the Elk Point Basin, created perhaps by the Meadow Lake Escarpment to the northwest in west central Saskatchewan and east central Alberta. Ashern deposition was terminated by a brief regression, represented by a breccia at the Ashern - Lower Winnipegosis contact in the Tableland area (Martindale and MacDonald, 1989).

Winnipegosis sedimentation was initiated during a Late Eifelian marine transgression (Fig. 3). The Winnipegosis Formation is divided into Lower and Upper Members. The Lower Member is composed of bituminous dolomitic mudstones and wackestones which vary little in thickness or composition in this area. These strata dip gently towards the centre of the basin and are interpreted to have been deposited under normal marine platform or ramp conditions on the Ashern Formation (Martindale and MacDonald, 1989).

The Upper Winnipegosis Member is Givetian in age and, in the Hitchcock area, consists of isolated buildups which grew in deeper water toward the centre of the Elk Point Basin (Fig. 2). In 7-19-3-8W2, the only well to core the Lower/Upper Winnipegosis contact in the Hitchcock area, there is no evidence of a bioconstructed foundation for the overlying buildup. A similar situation in the Tableland area led Martindale and MacDonald (1989) to conclude that buildups developed on structural, rather than stratigraphic, features.

Buildups at Hitchcock exceed 60 m in thickness and consist of grainstones, packstones, framestones and bindstones. The lower two-thirds are composed of codiacean algal and peloidal packstones and grainstones. The upper one third of the buildups consists of branched corals, chaetetids, blue-green and red algae, which locally form a "true" reef framework. Buildups are capped by laminated and contorted mudstones, deposited under subaerial conditions which terminated buildup deposition. In Saskatchewan the Upper Winnipegosis is entirely dolomitized, whereas in North Dakota, buildups may retain part of their original limestone lithology.

In basinal areas between buildups, the equivalent deposits consist of finely-laminated, dolomitic mudstones (these have previously been correlated with the Ratner Member in the Saskatoon area as defined by Reinson and Wardlaw (1972), but recent work by Kendall and co-workers has thrown doubt on this correlation - Kendall, *pers. comm.* 1991) (Fig. 3). In the 15-7-3-8W2M well, these laminites are 10 m thick and directly overlie the Lower Winnipegosis platform (Fig. 11). The laminites in the Tableland area are extensively dolomitized and can have high porosity and permeability (Martindale and MacDonald, 1989). Deposition of these laminites continued under increasingly evaporitic, albeit subaqueous, conditions following emergence of the reefs (Davies and Ludlam 1973). A highly bituminous mudstone, which may be equivalent to the informally named Brightholme member (Stoakes *et al.*, 1987) is commonly developed at the base of the laminites (Figs. 3, 11). Osadetz *et al.* (1990) have demonstrated that this unit, which is 1.5 m thick in the Hitchcock area (Fig. 11), is the source of most Winnipegosis oil in southeastern Saskatchewan.

The Presqu'ile Barrier in northwestern Alberta isolated the Elk Point Basin from the open ocean during a mid-Givetian sea level fall (Hallam, 1984). Increased hypersalinity resulted in precipitation of anhydrite, halite and potash salts of the Lower Prairie Evaporite (Jordan, 1967, 1968; Reinson and Wardlaw, 1972). In basinal areas between buildups, anhydrite varies from 3 to 5 m in thickness. Adjacent to the buildups, however, anhydrite thickens to greater than 20 m where it forms a "halo" surrounding the buildup (Fig. 3) (Bebout and Maiklem, 1973; Kendall, 1975; 1989; Martindale and MacDonald, 1989; Wardlaw and Reinson, 1971). Kendall (1989) has suggested that this anhydrite halo formed at springs on the reef surface when calcium-rich groundwaters mixed with basinal brines rich in sulphate to form gypsum. The anhydrite thins to 1-3 m over the tops of the buildups where it forms a seal for entrapped hydrocarbons (Fig. 3). Increasingly hypersaline conditions during the late Givetian produced the halites and potash salts of the Upper Prairie Evaporite.

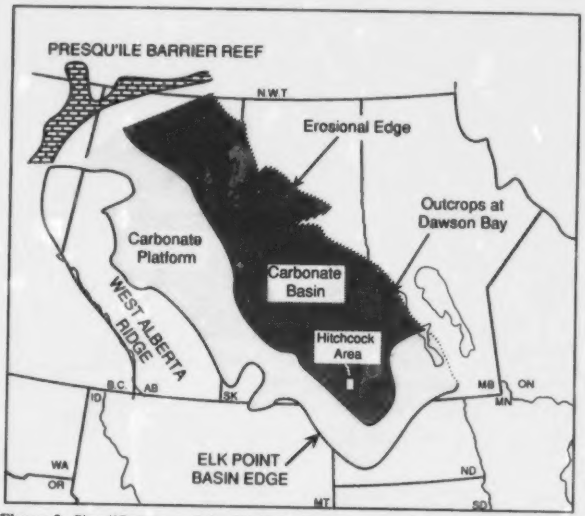


Figure 2: Simplified map of the Elk Point Basin and Presqu'ile Barrier during the early Givetian (modified from Williams, 1984; Grayston *et al.*, 1964; Martindale and MacDonald, 1989).

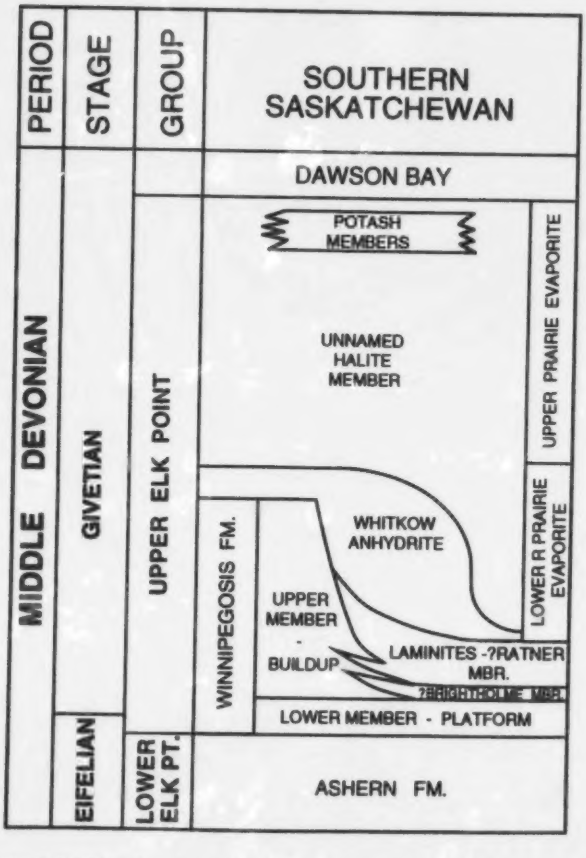


Figure 3: Middle Devonian stratigraphy in southeastern Saskatchewan (modified from Reinson and Wardlaw, 1972).

SEISMIC INTERPRETATION OF THE HITCHCOCK BUILDUPS

Seismic exploration is the primary tool for the detection of Winnipegosis buildups. The Hitchcock area was initially investigated with 2-D seismic lines spaced at 800 m (2,600 ft.). This was followed with a 11.5 km^2 (4.4 mile^2) 3-D seismic program. Dynamite was used as a source with line spacings of 400 m (1,312 ft.). Receiver lines were spaced at 300 m (984 ft.) and both shotpoints and receivers were spaced at 50 m (164 ft.) intervals. The resultant average subsurface coverage was approximately 2,000 percent in bins of 25 x 25 m (82 x 82 ft.). The processing sequence was standard for the area with particular emphasis placed on subsurface imaging using a full 3-D migration technique.

Upper Winnipegosis buildups in the Hitchcock area occur as chain-like complexes consisting of isolated reefs, conical in shape with steep sides and flat tops. To date, four wells (7-19-3-8, 15-19-3-8, 9-25-3-9 and 14-25-3-9W2M) have intersected the buildups which range from 57 to 63 m (187 to 207 ft.) in thickness and from 23 to 31 ha (57 to 77 acres) in area (Fig. 4).

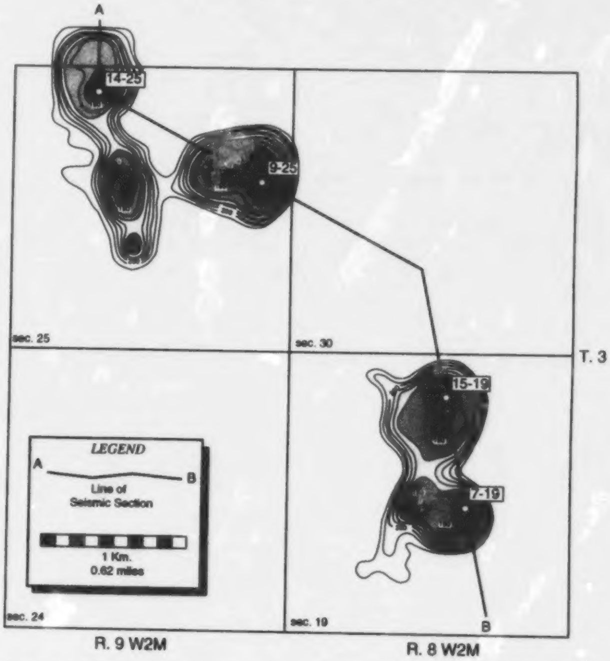


Figure 4: Torquay to Winnipegosis 3-D Isochron map (two millisecond contour interval) showing the location of wells penetrating the Hitchcock buildups and the location of the seismic line shown in Figure 5.

A composite 3-D seismic line (Fig. 5) through the four wells illustrates the strong acoustic impedance contrast (peak) representing the base of salt or Winnipegosis event at approximately 1.59 seconds two-way time (2,500 m / 8,200 ft.). The buildups themselves are seen as positive structural features with 20 - 25 milliseconds of relief. Negative structure on the Prairie Evaporite event, sub-Winnipegosis velocity pull-up and the replacement of the regional Winnipegosis reflection with an event representing the Ashern - Interlake contact are also characteristics associated with Winnipegosis buildups.

The "reef" anomaly is actually the seismic expression of the anhydrite halo which envelops the carbonate buildup. The dominant frequency at the Winnipegosis level is only 35 Hz., preventing reliable correlation of buildup interior reflectivity. A zero-phase synthetic seismogram from the 9-25-3-9W2M well demonstrates a strong tie between well data and seismic (Fig. 5). Buildups are also characterized by subtle paleotopographic highs or "crests" which may exhibit local amplitude variation. These crestal features are the primary drilling targets (Fig. 6).

Factors such as variable anhydrite thickness and limited vertical and spatial resolution are deterrents to detailed seismic character analysis. However, the accuracy of the 3-D migration process is undoubtedly the most important constraint on detailed structural and stratigraphic interpretation of the Winnipegosis seismic event.

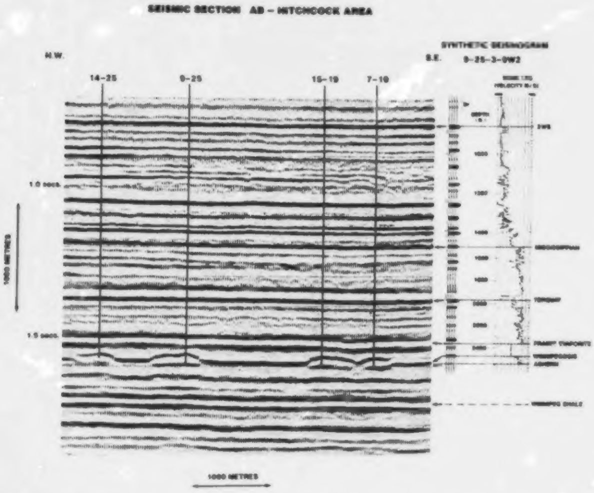


Figure 5: Composite seismic line through the Hitchcock wells and a synthetic seismogram for 9-25-3-9W2M.

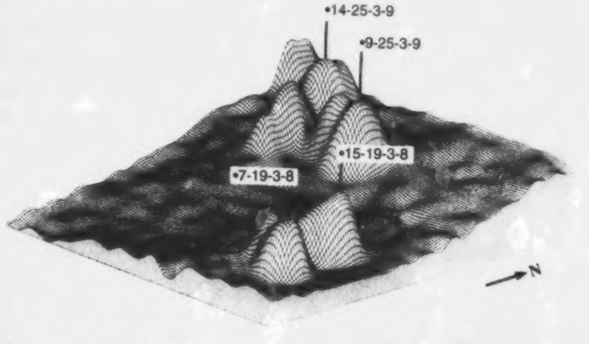


Figure 6: Isometric view of the Winnipegosis surface at Hitchcock derived from 3-D seismic data. View is from the southeast at an elevation of 20°. Vertical exaggeration 6:1.

METHODS OF INVESTIGATION

In the first three wells at Hitchcock (14-25, 9-25 and 15-19), cores were cut on penetration of the capping anhydrite. In 7-19 a single core was cut at the base of the buildup. Porosity, permeability, bulk density and oil saturation were analysed using full - diameter samples. On completion of the analyses, cores were slabbed and photographed and then selected samples were polished to reveal fine textural and faunal details. Polished slabs were examined by binocular microscope, described in detail (Figs. 7-10) and thin sections made of representative fabrics. In general, completion of the well was not undertaken until the slabbed core had been examined.

SEDIMENTOLOGY OF THE WINNIPEGOSIS BUILDUPS AT HITCHCOCK

A detailed core log incorporating petrophysical logs, drill stem test results, locations of perforations, lithological, sedimentological, faunal and diagenetic data and an interpretation of the depositional facies has been generated for each of the cored wells in the Hitchcock area (Figs. 7-10). Because each buildup has been penetrated by only one well, it is not possible to construct a cross section through a single buildup to show internal facies distribution. Instead, a composite facies model has been generated using the core and petrophysical log data from all of the wells at Hitchcock (Fig. 12). The relative location of each well on the buildup cannot be deduced from seismic data and this probably reflects the lack of distinct porosity zones associated with the various facies within the buildup. It may also reflect the rapid lateral changes in facies, described by Teare (1990) in the Winnipegosis of the Manitoba outcrop belt. Changes such as these, however, are beyond the current resolution limits of the seismic method at this depth.

LOWER WINNIPEGOSIS MEMBER

Crinoid Wackestone Facies

The Lower Winnipegosis platform has been drilled in four wells (15-7, 9-25, 7-19, 15-19) but has been cored only in the 7-19 well (Fig. 10). It varies from 11 to 12 m thick beneath the buildups and thins to 7 m at 15-7-3-8W2M (Fig. 11), away from the buildup complex. The Lower Winnipegosis - Ashern contact is clearly visible as an increase downward in the gamma ray log. The upper contact with the Upper Winnipegosis has been cored in 7-19 and is characterized by a minor increase downward in the gamma ray log response. A similar response and an associated decrease in the density log porosity can be seen at this contact in the uncored wells.

In the 7-19 core, platform strata are composed of very dark grey-black, nodular, bituminous dolomitic wackestones and mudstones with abundant crinoid fragments and local areas of anhydrite replacement (Fig. 14a). This facies is comparable to that described from the upper part of the platform in the Tableland area (Martindale and MacDonald, 1989). At Hitchcock, the platform thickens under the buildups. However, from the limited core data, it does not appear that the platform forms a bioconstructed foundation for the overlying reef. Reservoir quality of this member is generally poor.

The Lower Winnipegosis platform sediments represent deposition in an open marine, deep water setting of normal salinity, probably on a shelf or gently sloping ramp after the initial transgression of the Elk Point sea (Fig. 13).

UPPER WINNIPEGOSIS MEMBER

Peloid/Codiacean Algal Grainstone Facies

This facies, which is estimated to be 45 m thick in 9-25 and 15-19 forms the lower two-thirds of the Hitchcock buildups. It directly overlies the Lower Winnipegosis platform and is overlain

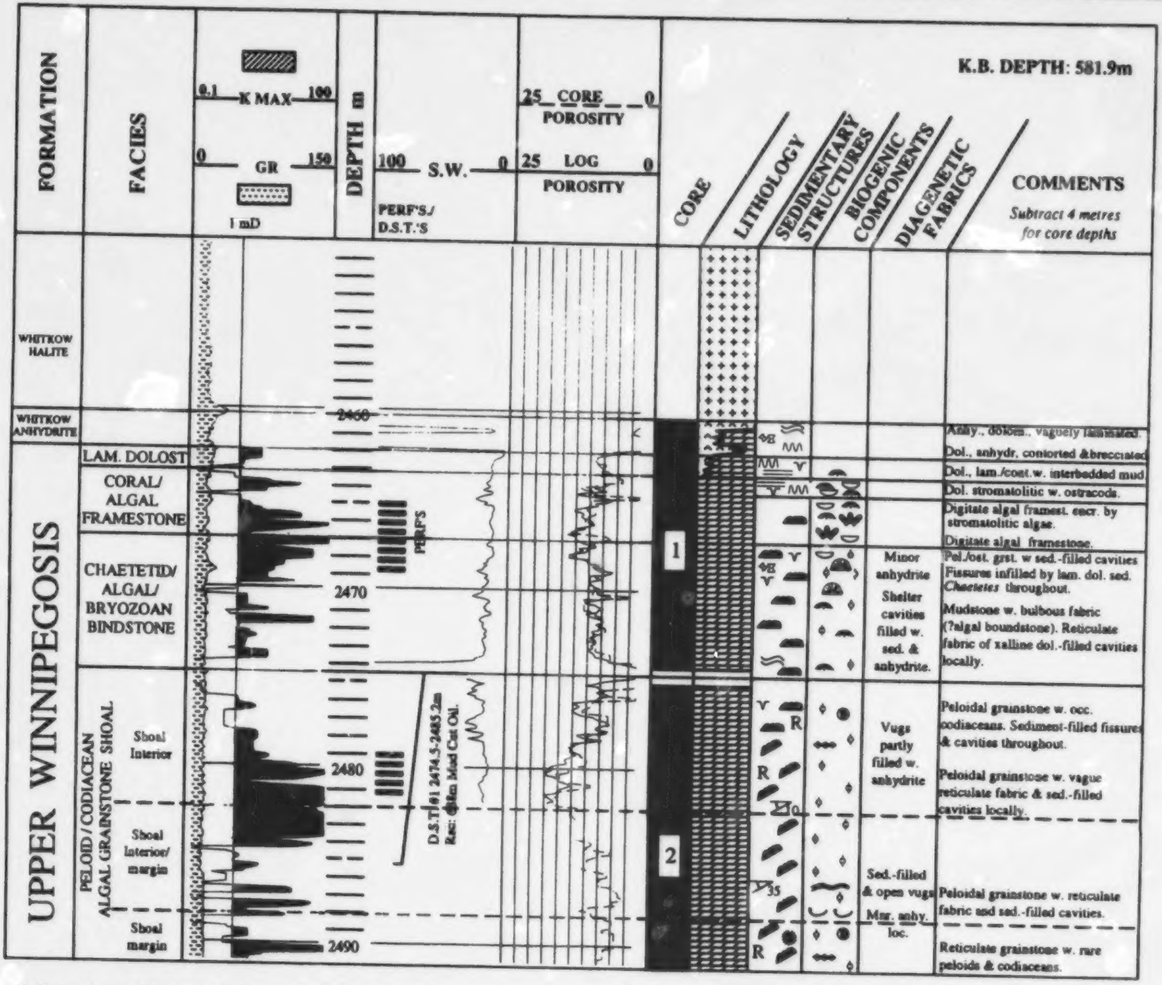

FORMATION
FACIES
0.1 K MAX 100
0 GR 150
1 mD
DEPTH m
100 S.W. 0
PERF'S./ D.S.T.'S
25 CORE POROSITY 0
25 LOG POROSITY 0
CORE
LITHOLOGY
SEDIMENTARY STRUCTURES
BIOGENIC COMPONENTS
DIAGENETIC FABRICS
K.B. DEPTH: 581.9m
COMMENTS
Subtract 4 metres for core depths
WHITKOW HALITE
WHITKOW ANHYDRITE
UPPER WINNIPEGOSIS
LAM. DOLOST
CORAL/ ALGAL FRAMESTONE
CHAETETID/ ALGAL/ BRYOZOAN BINDSTONE
PELOID / CODIACEAN ALGAL GRAINSTONE SHOAL
Shoal Interior
Shoal Interior/ margin
Shoal margin
2460
2470
2480
2490
PERFS
D.S.T.#1 2474.5-2485.2m
Rec: 488m Mud Cut Oil.
1
2
R
35
Minor anhydrite
Shelter cavities filled w. sed. & anhydrite.
Vugs partly filled w. anhydrite
Sed.-filled & open vugs
Mnr. anhy. loc.
Anhy., dolom., vaguely laminated.
Dol., anhydr, contorted &brecciated
Dol., lam./cont.w. interbedded mud
Dol. stromatolitic w. ostracods.
Digitate algal framest. encr. by stromatolitic algae.
Digitate algal framestone.
Pel./ost. grst. w sed.-filled cavities
Fissures infilled by lam. dol. sed.
Chaetetes throughout.
Mudstone w. bulbous fabric (?algal boundstone). Reticulate fabric of xalline dol.-filled cavities locally.
Peloidal grainstone w. occ. codiaceans. Sediment-filled fissures & cavities throughout.
Peloidal grainstone w. vague reticulate fabric & sed.-filled cavities locally.
Peloidal grainstone w. reticulate fabric and sed.-filled cavities.
Reticulate grainstone w. rare peloids & codiaceans.

Figure 7: Composite well log for 14-25-3-9W2M illustrating core lithology, sedimentary structures, biogenic components, test and completion intervals and petrophysical log data (GR - gamma ray, K Max - maximum analysed permeability, S.W. - water saturation). Calculated log porosity is derived from the density log using a grain density of 2840Kg/cubic metre. Symbols are explained in Fig. 12.

by the chaetetid/algal/bryozoan bindstone facies (Fig. 12). The grainstone facies is characterized by codiacean algal fragments and undifferentiated peloids, as well as scattered stromatoporoids and corals. These form grainstones and packstones which may possess steep dips locally as a result of early cement precipitation. Variations in fauna, and depositional and diagenetic fabrics allow the grainstone facies to be divided into three subfacies: shoal margin, shoal interior/margin and shoal interior.

The shoal margin subfacies is present in the lower part of the 15-19 and 14-25 cores. It achieves a maximum thickness of only 6 m here but is interpreted, on the basis of similar facies in the Tableland area, to occur throughout the uncored, lower part of the Hitchcock buildups and to directly overlie the Lower Winnipegosis platform (Figs. 7-9, 12). This subfacies comprises codiacean green algal fragments and scattered peloids in a recrystallized muddy matrix characterized by steep dips and an abundance of fibrous cement-lined and sediment-filled cavities (Figs. 14b, 14c, 14d). Because of this, permeabilities are highly variable.

The shoal interior/margin subfacies, which ranges from 3 to 9 m in thickness, is present in all of the cored wells at Hitchcock, where it overlies the shoal margin facies (Fig. 12). Biogenic components and depositional and diagenetic fabrics are similar to those of the margin facies. However, both peloids and codiacean algal fragments are more abundant and there is a decrease in the number of steeply dipping cavities with fibrous cement linings (Figs. 14e, 14f, 15a). A reticulate fabric, composed of millimetre to centimetre - sized cavities which are lined by equant dolomite cement or filled with reef-derived peloidal sediment and crystalline dolomite, is common throughout this subfacies and occurs locally in adjacent subfacies (Figs. 15b, 15c). Reservoir quality in this subfacies is highly variable.

The shoal interior subfacies overlies the shoal interior/margin subfacies and varies from 5 to 17 m thick in cored wells and may be up to 37 m thick in 7-19 (Fig. 12). This subfacies is dominated by both large and small codiacean algae towards the base (*e.g.*, 7-19) and by peloids near the top (Figs. 15d, 15e, 15f). Fenestrate bryozoa, chaetetids, stromatolitic algae, thamnoporid corals, crinoids, brachiopods and bulbous and hemispherical stromatoporoids are common locally, especially in 9-25. The reticulate fabric is common throughout but steeply-dipping cavities and fibrous cements are absent. Because of this, the shoal interior subfacies generally has better reservoir characteristics than the two underlying subfacies.

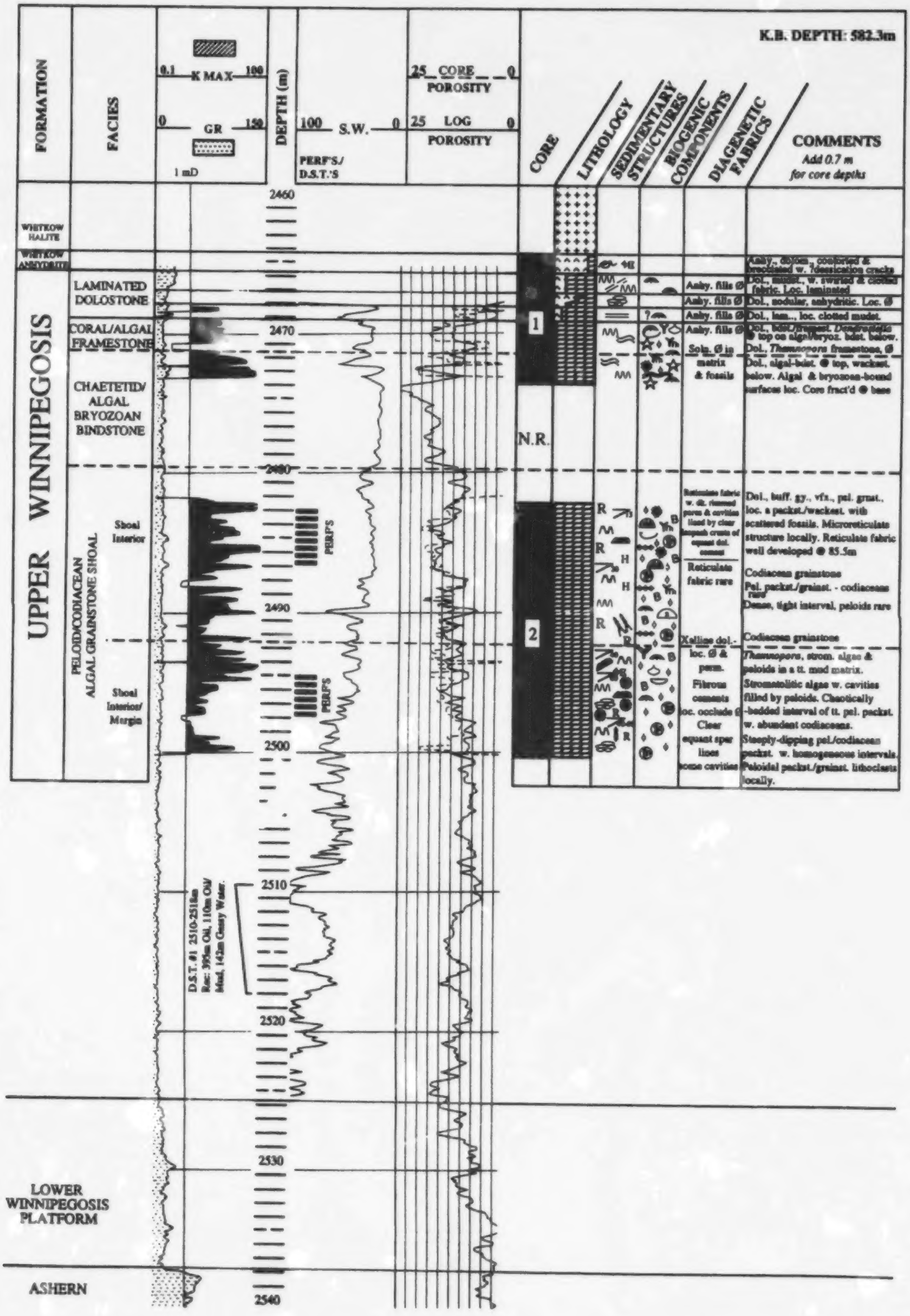


Figure 8: Composite well log of Scurry et al Hitchcock 9-25-3-9W2M

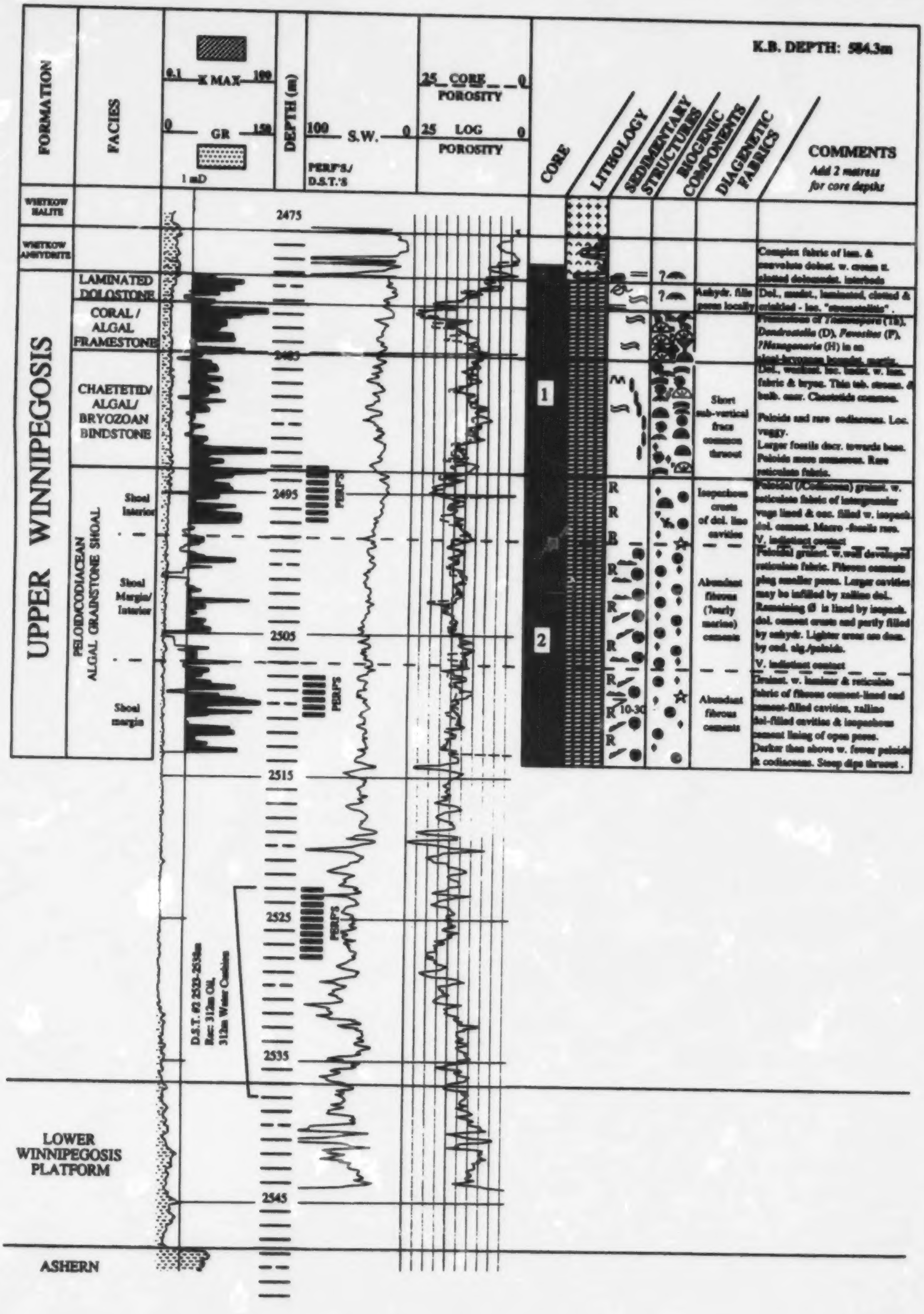
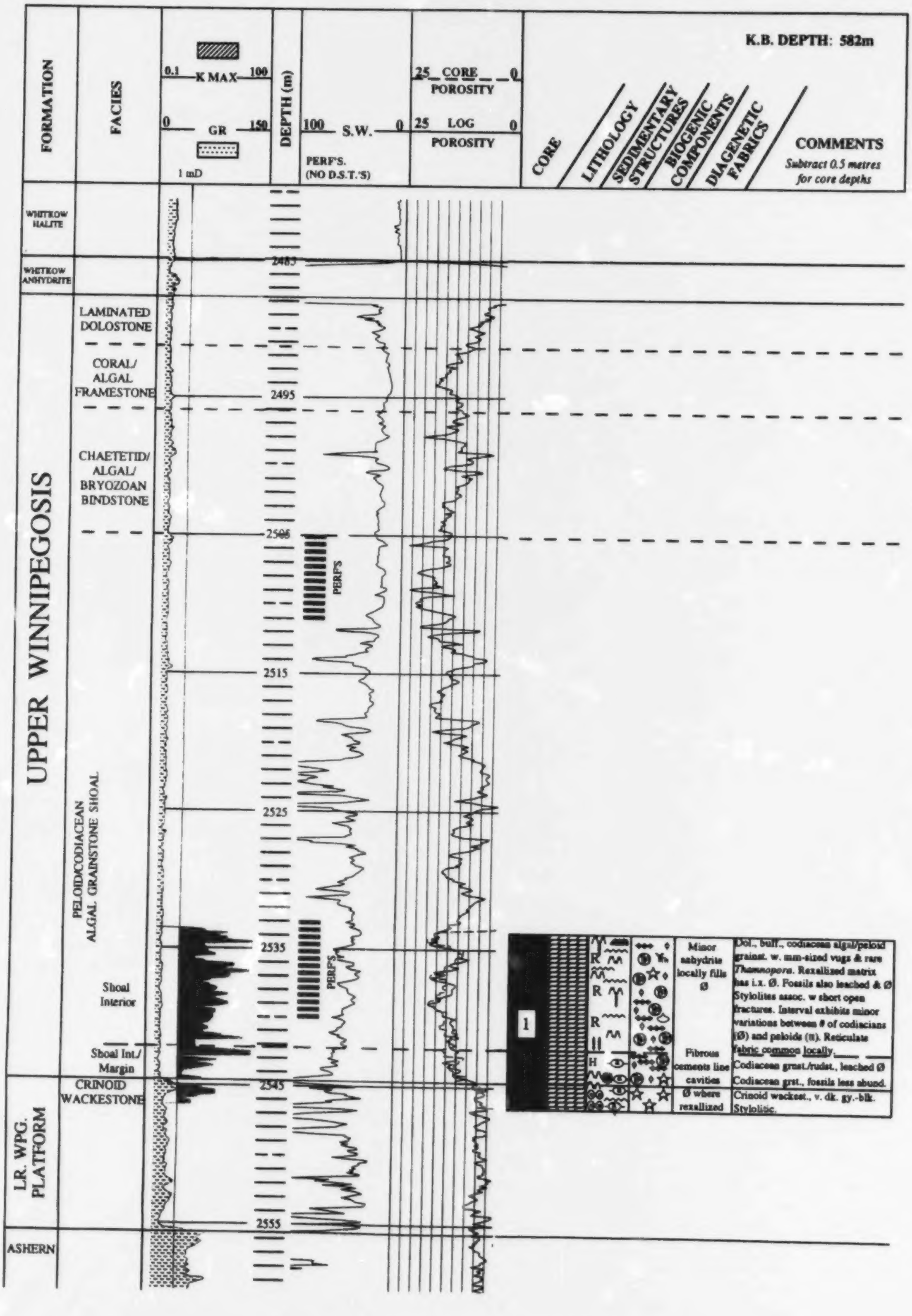


Figure 9: Composite well log of Scurry et al Hitchcock 15-19-3-8W2M



Minor anhydrite locally fills Ø

Fibrous cements line cavities

Ø where rexallized

Dol., buff., codiacean algal/peloid grainst. w. mm-sized vugs & rare *Thamnopora*. Rexallized matrix has i.x. Ø. Fossils also leached & Ø Stylolites assoc. w short open fractures. Interval exhibits minor variations between # of codiacians (Ø) and peloids (π). Reticulate fabric common locally.

Codiacean grnst./rudst., leached Ø

Codiacean grst., fossils less abund.

Crinoid wackest., v. dk. gy.-blk. Stylolitic.

Figure 10: Composite well log of Scurry Hitchcock 7-19-3-8W2M

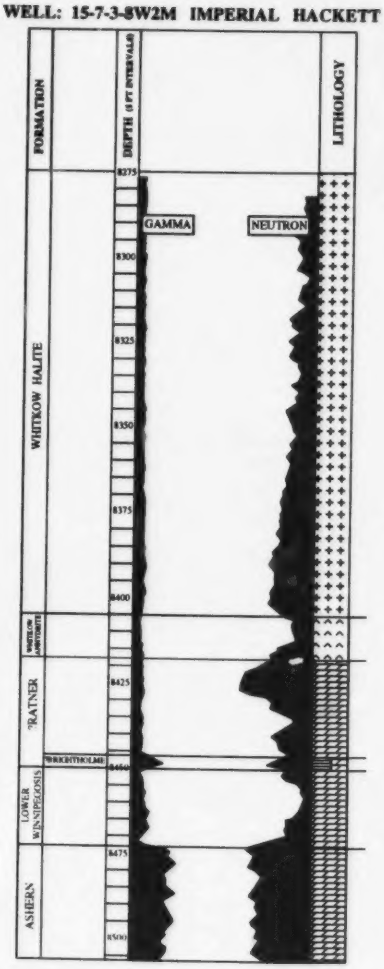


Figure 11: Simplified well log for the uncored well 15-7-3-8W2M. Lithologies are interpreted from log data.

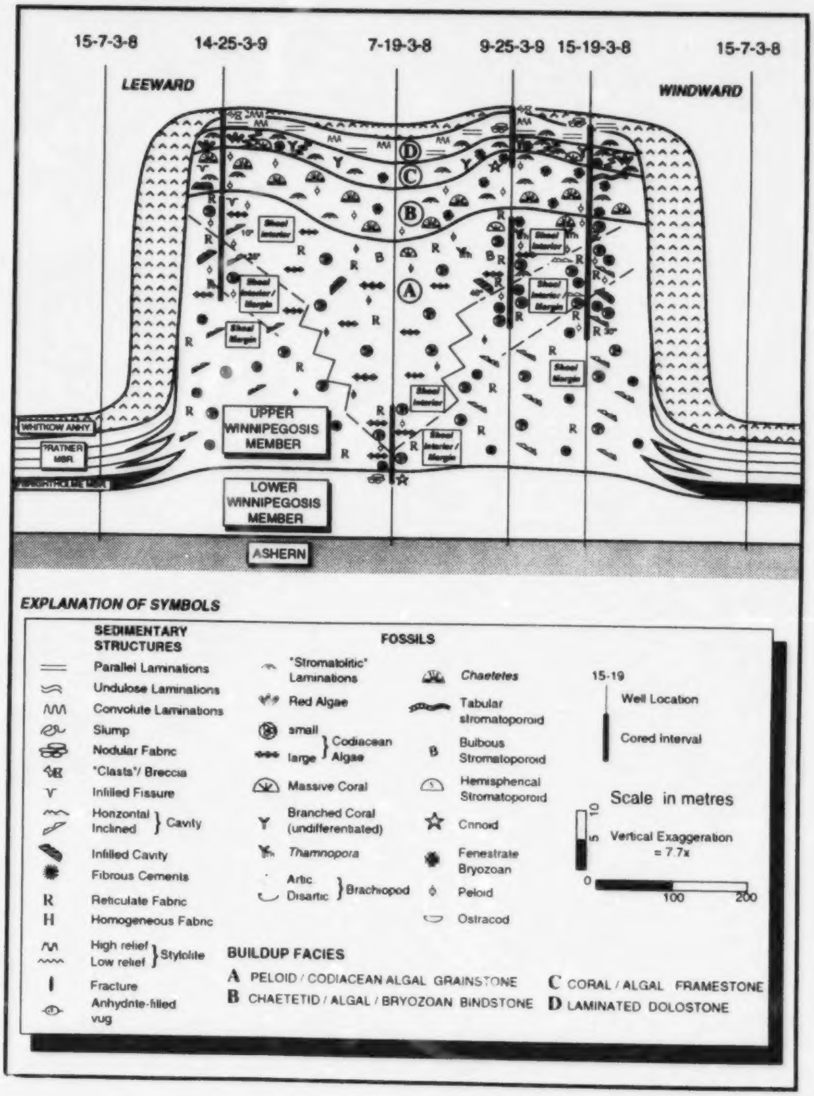


Figure 12: Facies model, derived from the reefs at Hitchcock, showing the relative location of wells on a hypothetical buildup.

The Peloid/Codiacean Algal Grainstone Shoal facies probably developed in moderately shallow water of normal salinity. Although codiacean algae are present in all three subfacies, those of the shoal interior are the largest in size and the most abundant. The codiacean algae (genus *Litanaia* and *Paralitanaia* Maslov 1956), currently under review as Udoteacean algae (B. Mamet, *pers. comm.*, 1990), probably grew as upright colonies, which disarticulated upon death to form grainstone and packstone deposits. Further colonization and disarticulation of the algae formed shoals which eventually coalesced due to lateral and vertical accretion (Fig. 13). Seaward margins were stabilized by early marine, fibrous cements, precipitated from circulating sea water in a manner similar to that described by James and Ginsburg (1979) in Belize and by Marshall and Davies (1981) in the Great Barrier Reef. In this way, slopes of 30° to 40° could be maintained.

Support for this interpretation is provided by recent studies in the Caribbean (Hine *et al.*, 1988), Java Sea (Roberts *et al.*, 1988) and on the Great Barrier Reef (Marshall and Davies, 1988), where buildups, composed of the codiacean alga *Halimeda*, occur in both shallow and deep marine environments. These buildups have steeply-dipping margins, internal cavities, high primary porosities and some degree of wave resistance. Seismic evidence shows them to be composed of a number of smaller, coalesced shoals, similar in shape and size to those identified in the Winnipegosis Formation.

Shoal interiors are dominated by peloids, with only minor codiaceans and other organisms. In these quieter water areas, peloids may have been generated by intense reworking of codiacean and other biogenic debris.

The superposition of shoal interior on shoal margin subfacies illustrates the outward progradation of the buildups with time. The thickness of each of these subfacies allows cored wells to be placed in their relative position in the hypothetical facies model (Fig.12). The shoal facies of 7-19 is thinner than the equivalent facies in more marginal wells. This may be due to greater compaction of the relatively uncemented shoal interior sediments. Anderson and Franseen (1991) have shown that Winnipegosis reefs in the Oxbow area (Twp. 2 Rge. 3 W2M) have raised, cemented rims and depressed interiors. They have suggested that the greatest porosity will be found in the reef rim despite early cementation. Both the Hitchcock and Tableland buildups, however, demonstrate that the reef interior facies has the best reservoir characteristics.

Chaetetid/Algal/Bryozoan Bindstone Facies

This facies overlies the Peloid/Codiacean Algal Grainstone facies and has been cored in three of the four Hitchcock wells. It ranges from 7 to 8 m thick in 14-25 and 15-19 and probably has a similar thickness in 9-25 and 7-19. Although the lower boundary is gradational, the upper boundary with the Coral/Algal Framestone is generally better defined. Stromatolitic algae and fenestrate bryozoa locally form bindstones in 9-25 and 15-19 and chaetetids and stromatolitic algae form bindstones and framestones in 14-25 (Figs. 16a-16c). The matrix is dominated by peloids with rare ostracods and crinoids. Codiacean green algae are rare or absent. This facies is characteristically more massive and homogeneous than those below. Although reticulate fabrics, fibrous cements and steeply dipping cavities are rare, small shelter cavities within bindstones are common locally and are commonly geopetally filled with reef-derived peloidal sediment. Larger cavities and fissures contain a later fill of grey laminated microcrystalline dolostone sediment which is devoid of biogenic material (Figs. 16b, 16c).

Porosity and permeability are lower than in the underlying shoal facies but in general permeabilities are more uniform, reflecting the greater homogeneity of this reservoir.

Faunal content and sedimentary styles in this facies are markedly different from those in the underlying shoal facies. Steep dips are rare and the more homogeneous character of the sediment suggests a reef interior environment of deposition. The abundance of fenestrate bryozoa and stromatolitic algae also suggest deposition in quiet waters in the photic zone (Fig. 13).

Larger cavities and sub-vertical fissures (Fig. 16b) are interpreted as karst features which formed during subaerial exposure of the buildup and which have subsequently been filled by vadose silt (Dunham 1969, Maiklem 1971). Similar features are evident in the nearby Tableland buildup (Martindale and MacDonald, 1989).

Coral/Algal Framestone Facies

This facies has been cored in three of the four wells at Hitchcock where it varies from 2.9 to 3.7 m thick. Biogenic components form a true "reef" framework. In 14-25, the red alga *Solenopora* occurs as exceptionally well-preserved, upright, bushy colonies, many of which are encrusted by *Sphaerocodium* (Figs. 16d, 16e). Bulbous frame-building stromatolitic algae overlie this interval. The equivalent facies in 15-19 and 9-25 is dominated by the branching corals *Thamnopora* and *Dendrostella* (which may locally be encrusted by stromatolitic algae), by more massive corals such as *Favosites* and ?*Hexagonaria* (Fig. 16f) and by bulbous stromatolitic algal colonies (9-25 only). In all three wells, the matrix comprises crinoid fragments, ostracods and peloids (possibly of algal origin) which are locally bound by stromatolitic algae and fenestrate bryozoa.

Solution vugs and shelter cavities are present locally. The quality of the reservoir in this facies is generally good, especially where corals and digitate algae are leached. Larger voids and cavities of probable karst origin are present in 14-25 and are filled with laminated dolostone (vadose silt) which reduces porosity and permeability.

This facies represents a true reef framework. Counterparts of all of the constituents occur today on reefs growing in warm, shallow waters of normal salinity. A certain degree of reef zonation is apparent at both Hitchcock and Tableland (Martindale and MacDonald, 1989) and in the Dawson Bay region of the Manitoba outcrop (Teare, 1990; Fig. 2). In the hypothetical model, the solenopores are assigned a shallow water reef margin or crest position (Fig. 12). This distribution is in accord with that of modern coralline algae, probable descendants of the solenopores, which form wave resistant platforms and ridges on the windward margins of Caribbean and Pacific reefs (Adey, 1975; Adey and Burke, 1976; Wray, 1977). Massive and branching corals are assumed to occupy a more sheltered position behind the algal crest. The presence of stromatolitic algae in the upper part of this facies in 14-25 suggests that later stages of development may have occurred on a raised reef rim at or close to sea level (Fig.13).

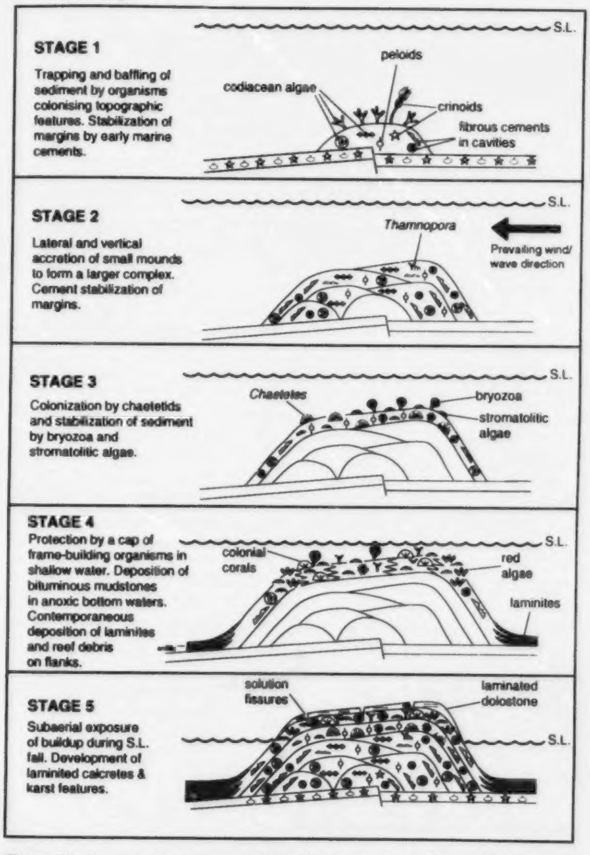
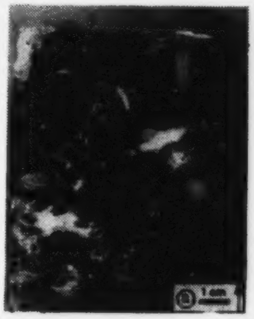
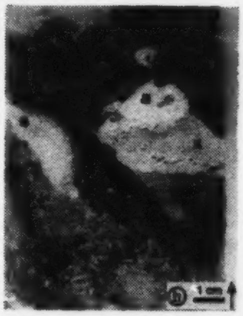
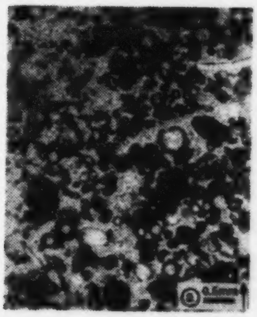
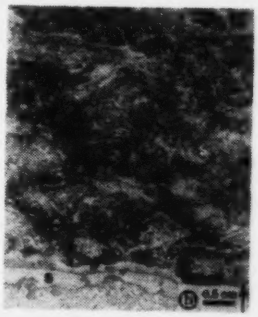

S.L.
STAGE 1
Trapping and baffling of sediment by organisms colonising topographic features. Stabilization of margins by early marine cements.
peloids
codiacean algae
crinoids
fibrous cements in cavities
S.L.
STAGE 2
Lateral and vertical accretion of small mounds to form a larger complex. Cement stabilization of margins.
Thamnopora
Prevailing wind/ wave direction
S.L.
STAGE 3
Colonization by chaetetids and stabilization of sediment by bryozoa and stromatolitic algae.
Chaetetes
bryozoa
stromatolitic algae
STAGE 4
Protection by a cap of frame-building organisms in shallow water. Deposition of bituminous mudstones in anoxic bottom waters. Contemporaneous deposition of laminites and reef debris on flanks.
S.L.
colonial corals
red algae
laminites
solution fissures
laminated dolostone
STAGE 5
Subaerial exposure of buildup during S.L. fall. Development of laminited calcretes & karst features.
S.L.

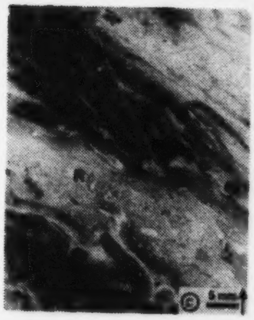
Figure 13: Stages in the development of a Winnipegosis buildup at Hitchcock and adjacent areas.

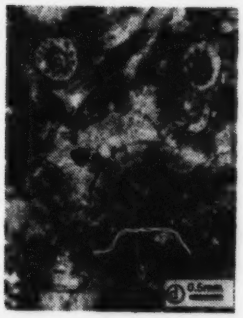


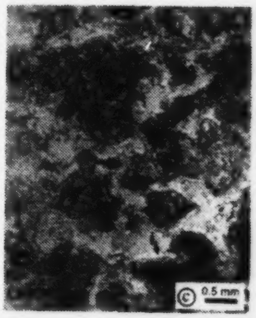


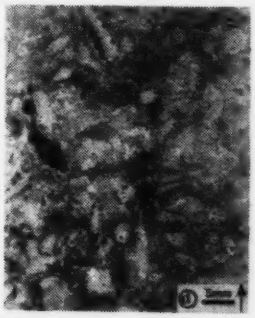


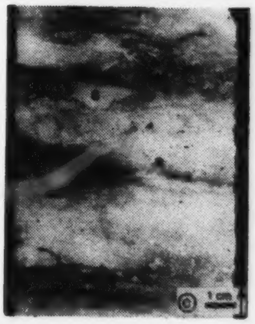


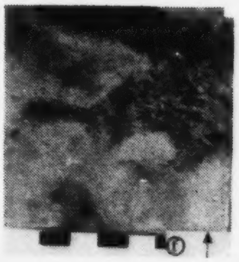












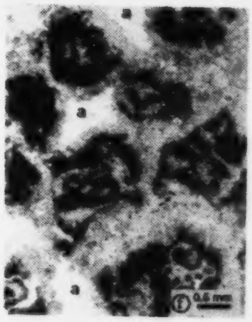


Figure 14a: 7-19-3-8W2M 2545.4 m. Crinoid Wackestone Facies. Dark, bituminous, anhydritic, crinoidal mudstone of the Lower Winnipegosis platform.

Figure 14b: 14-25-3-9W2M 2486.7 m. Shoal Margin subfacies. Reticulate-patterned (R) peloid/codiacean mudstone with a solitary coral (c) and stromatoporoid fragment (s). The vague 30° dip in this core is emphasized by the margins of the cavity. This has been infilled by peloidal packstone (p) containing a gastropod (arrowed), crystalline dolomudstone (x) (probably vadose silt) and anhydrite (a).

Figure 14c: 15-19-3-8W2M 2512.35 m. Shoal Margin subfacies. Reticulate-patterned (R) peloid/codiacean algal wackestone characterized by steeply-dipping, fibrous cement-lined primary shelter and growth cavities (c).

Figure 14d: 15-19-3-8W2M 2515.9 m. Shoal Margin subfacies. Thin section photomicrograph of a wackestone consisting of codiacean algae (c) in a recrystallised ?peloidal matrix. Pores are lined and almost filled by banded, fibrous isopachous cement (f).

Figure 14e: 15-19-3-8W2M 2500.5 m. Shoal Interior/Margin subfacies. Reticulate-patterned, peloid/codiacean algal packstone with numerous flat-lying, lenticular cavities. Most are filled with peloidal wackestone sediment (s) but locally, late-stage anhydrite (a) occludes remaining porosity.

Figure 14f: 14-25-3-9W2M 2485.25 m. Shoal Interior/Margin subfacies. Nodular fabric of homogeneous peloidal wackestones (p) within a "matrix" of reticulate patterned, codiacean/peloid packestone (c) with fibrous cements locally lining pores. Scale divisions: 1 cm.

Figure 15a: 14-25-3-9W2M 2476.65 m. Shoal Interior/Margin subfacies. Thin section photomicrograph of a wackestone (locally a packstone) of poorly-preserved irregular peloids, shell fragments, ?ostracods and various algal cysts (p - *Parathurammina*, b - *Biospheras*) in a dolomitized, microcrystalline matrix.

Figure 15b: 15-19-3-8W2M 2500.5 m. Shoal Interior/Margin subfacies. Pronounced reticulate fabric of numerous flat-lying, interconnected cavities in a peloid (codiacean algal) packstone matrix. Larger cavities are partly filled by peloidal sediment(s). Crystalline dolomite fills the remaining, particularly smaller cavities. Open pores result from later dissolution and are lined by equant cement crusts.

Figure 15c: 9-25-3-9W2M 2485.5 m. Shoal Interior subfacies. Thin section photomicrograph of reticulate fabric consisting of poorly-preserved codiacean algal fragments (c) in a porous dolomudstone matrix. Pores are lined by isopachous crusts of equant dolomite cement (i).

Figure 15d: 7-19-3-8W2M 2538.85 m. Shoal Interior subfacies. Wackestone of larger codiaceans (?*Paralitanaia*) in a peloidal packstone matrix. Most of the codiaceans have undergone some dissolution of the outer cortical region.

Figure 15e: 14-25-3-9W2M 2478.4 m. Shoal Interior subfacies. Homogeneous peloid grainstone with scattered mm-sized pores and rare, small codiacean algal fragments.

Figure 15f: 7-19-3-8W2M 2541.81 m. Shoal Interior subfacies. Moderately well-preserved codiacean algal fragments "float" in a porous dolostone matrix. Anhydrite (a) occludes porosity locally.

Laminated Dolostone Facies

This facies, which can be readily identified in both cores and petrophysical logs, varies from 1.2 to 3.6 m in thickness. It is typically a flat-lying, finely (millimetre-scale) laminated dolostones with interbedded (millimetre-sized) fenestrae. The fenestrae are bounded by a dark, organic? film and are usually lined with an isopachous equant dolomite cement crust which may, together with anhydrite, completely fill the pore (Figs. 17a, 17b). Locally, the laminated dolostone is contorted and may be interbedded with nodular anhydrite. Pisolites similar to those described by Wardlaw and Reinson (1971) are present in 15-19 (Fig. 17c). Horizontal and subvertical sediment-filled fissures disrupt the laminated crusts at a number of locations in 14-25 (Fig. 17d). Except for laminated fabrics which may be of microbial origin, no fossils were recognized in this facies.

Previous interpretations of the Winnipegosis Formation, particularly in the Manitoba outcrop belt (McCabe and Barchyn, 1982), have suggested that the laminar dolostones of the Upper Winnipegosis are stromatolitic. While undoubtedly similar in character to cryptalgal fabrics illustrated by Flugel (1982), the laminar dolostones are also similar to calcretes illustrated by Wright *et al.* (1988), to calcretes in the Tableland area (Martindale and MacDonald, 1990) and to calcretes examined by the author (W.M.) on Grand Cayman, West Indies. Teare (1990) also concluded that the upper 4 m of laminated Winnipegosis dolostone seen in outcrop at Dawson Bay, Manitoba represent calcrete deposits.

Calcretes imply that buildups were subjected to subaerial exposure. Vadose silt which formed during this period percolated into both karst-formed and primary vugs and cavities. Other vadose fabrics such as pendant cements, recognized elsewhere (Martindale and MacDonald, 1989), were not observed in the Hitchcock wells.

DIAGENESIS OF THE WINNIPEGOSIS FORMATION

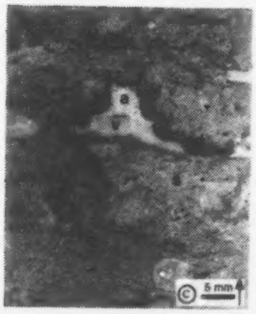
As many of the diagenetic fabrics seen in the Hitchcock buildups have already been described from the Tableland area (Martindale and MacDonald, 1989), the following discussion will review the influence of some of the more significant diagenetic fabrics on buildup growth, morphology and reservoir characteristics. Significant diagenesis of the Winnipegosis began with precipitation of fibrous cements within primary cavities between codiacean fragments (Figs. 14d and 17e). As the shoals lack any evidence of organic stabilization of their steep flanks, it has to be concluded that these early marine cements were responsible for the preservation of high angle slopes, which permitted vertical accretion and the formation of significant topographic features. The origin of different cavity types within Winnipegosis buildups is discussed in detail by MacDonald (1989), MacDonald and Jones (1989) and Martindale and MacDonald (1989). Both reticulate cavities, which formed between cemented crusts in peloid and codiacean grainstones, and flat-bottomed cavities which are more ubiquitous, are sites of precipitation of marine fibrous cements and of contemporaneous internal sediment (Figs. 14b, 15b, 16c). Vadose silt and anhydrite commonly fill any remaining porosity (Figs. 14b,16c). Vertical and subvertical fissures are present in the upper 13 m of the 14-25 buildup (Figs. 16b, 17d). These karst features are filled by laminated crystalline dolostone similar to that found in primary cavities. Secondary cavities such as these are common in Tableland and other neighbouring buildups where they form an interconnected network which may extend 40 m below the buildup surface (*e.g.*, 14-35-4-10W2M, Martindale and MacDonald, 1989).

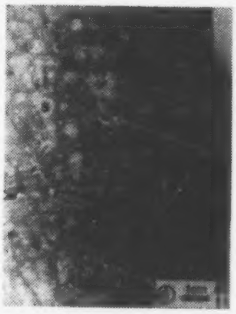
Dolomitization of the Winnipegosis buildups at Hitchcock has not been investigated in detail. Stanford (1989) described the nature and origin of dolomites in this general area and the evidence from Hitchcock generally supports her interpretation. The limestones in Winnipegosis buildups at Hitchcock are replaced by early, pervasive dolomite. Stanford has shown these dolomites to be non-stoichiometric and slightly enriched in magnesium. These factors led her to conclude that dolomitization occurred before deposition of the Prairie Evaporite salt by means of seepage reflux of concentrated marine brines. These brines were depleted in sulphate and relatively enriched in magnesium as a result of gypsum precipitation in surrounding basinal areas. At Tableland, there is some preservation of original limestone in the Lower Winnipegosis platform. Limestone has not been recognized in cores at Hitchcock but areas of low porosity in the Lower Winnipegosis platform at 9-25 (Fig. 8) may have resulted from incomplete dolomitization.

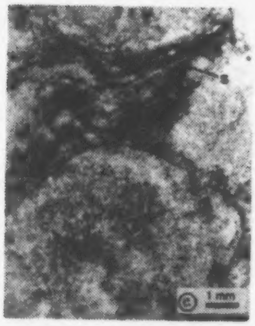
Early, pervasive dolomitization preserved much of the primary depositional fabric of the buildups. However, later recrystallization and dissolution (Stanford, 1989) locally destroyed finer cement fabrics and fossil microstructures and generated considerable intercrystalline and vuggy intrafossil porosity (Figs. 16e, 17f). The small cavities so formed are lined and partially occluded by an isopachous crust of equant dolomite (Figs. 15c, 17b, 17e), formed in the shallow burial environment (MacDonald, 1989; Martindale and MacDonald, 1989).

RESERVOIR QUALITY

The Upper Winnipegosis reservoir at Hitchcock is, in general, a heterogeneous system with highly variable porosity and permeability (maximum analysed porosity 20 percent, average porosity 11 percent, average permeability 3 mD). Reservoir characteristics are determined by primary depositional fabrics and diagenetic events which have modified primary porosity. The best reservoir is encountered in the shoal interior and chaetetid/algal/bryozoan bindstone facies where primary intergranular porosity has been enhanced by dolomitization. In contrast, precipitation of early marine cements may significantly reduce the high primary porosity of the shoal margin subfacies. Anhydrite generally reduces porosity in the laminated dolostone facies. Infilled vadose cavities and fissures form an interconnected, randomly distributed network throughout the upper porous parts of the buildup and can locally reduce vertical and horizontal permeability. In the lower parts of the buildup, primary pores and cavities which have escaped cement and silt infill have, where connected, excellent vertical permeability. This, however, is a property that may permit rapid ingress (coning) of water from the underlying aquifer. Production problems such as this are present in the Tableland field and in other wells in this area.







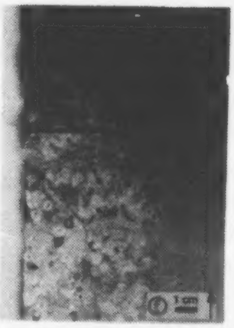


Figure 16a: 9-25-3-9W2M 2472.5 m. Chaetetid/Algal/Bryozoan Bindstone facies. Porous peloid packstone baffled and/or bound by numerous fenestrate bryozoa (b) Late-stage dissolution of corals (c) and matrix has generated secondary vuggy porosity.

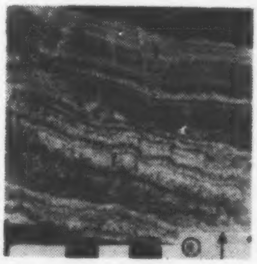
Figure 16b: 14-25-3-9W2M 2463.5 m. Chaetetid/Algal/Bryozoan Bindstone facies. Vadose silt (s), composed of tight, laminated microcrystalline dolostone sediment with no obvious fossils, fills shelter cavities and a broad solution fissure in a chaetetid framestone (c).

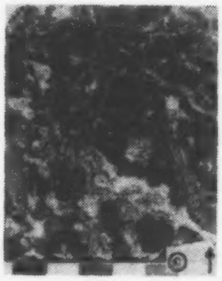
Figure 16c: 14-25-3-9W2M 2469.4 m. Chaetetid/Algal/Bryozoan Bindstone facies. Prominent shelter cavity in an ?algal-bound peloid grainstone. The cavity is lined by botryoidal algal colonies (b) and is partly filled by peloid grainstone sediment (s) and vadose silt (v). Anhydrite (a) fills the remaining pore space.

Figure 16d: 14-25-3-9W2M 2462.0 m. Coral/Algal Framestone facies. Framestone formed from digitate colonies of the red alga *Solenopora* (s).

Figure 16e: 14-25-3-9W2M 2462.15 m. Coral/Algal Framestone facies. Thin section photomicrograph of *Solenopora* encrusted by laminar algae, possibly *Sphaerocodium* (s). Later recrystallization and dissolution has destroyed much of the cell structure (c) of *Solenopora*, creating moderate intercrystalline porosity.

Figure 16f: 15-19-3-8W2M 2483.1 m. Coral/Algal Framestone facies. Framestone of *Dendrostella* (d) and *Thamnopora* (t) in a muddy, crinoidal wackestone matrix. This interval is abruptly overlain by dark mudstones of the Laminated Dolostone facies.





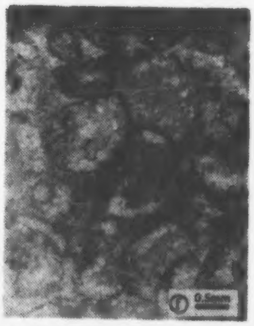


Figure 17a: 9-25-3-9W2M 2469.1 m. Laminated Dolostone facies. Finely laminated (l) and fenestral-laminated (f) dolostone, interpreted as a calcrete formed on the buildup during subaerial exposure. Scale divisions: 1 cm.

Figure 17b: 9-25-3-9W2M 2469.1 m. Laminated Dolostone facies. Thin section photomicrograph of the fenestral-laminated dolostones illustrated in Fig. 17a. Fenestrae are bounded by a dark, ?organic film and may by completely filled by late-stage equant dolomite. Elsewhere, dolomite forms an isopachous crust 1 crystal thick. Although porous, these intervals lack significant permeability.

Figure 17c: 15-19-3-8W2M 2482.3 m. Laminated Dolostone facies. Compound pisoliths occur in an interval 8 cm thick within the laminated dolostone facies. Pisoliths exhibit multiple encrustation and local concentric structure. They may represent storm reworking of the calcrete crust or they may be vadose pisolites developed under subaerial conditions. Scale divisions: 1 cm.

Figure 17d: 14-25-3-9W2M 2458.8 m. Laminated Dolostone facies. Laminar dolostones disrupted by two generations of solution fissures (1, 2). Fissures and irregular cavities are filled by crudely laminated vadose silt.

Figure 17e: 15-19-3-8W2M 2513.15 m. Shoal Margin subfacies. Cavities within a codiacean algal wackestone/floatstone are lined and locally filled by a thick crust of fibrous cement (f). This first generation marine cement is subsequently overgrown by a later phase of more equant, isopachous dolomite of shallow burial origin (i).

Figure 17f: 14-25-3-9W2M 2462.15 m. Coral/Algal Framestone facies. *Solenopora* colonies preserved as micritized rims. Dissolution and recrystallisation have removed the original cellular fabric and created considerable intrafossil porosity.

COMPLETION PROCEDURES AT HITCHCOCK

Completion practices in the Winnipegosis at Hitchcock undergo constant revision to improve formation inflow and lessen initial completion formation damage. The first completion at 14-25 was carried out conventionally using overbalanced wireline perforating. The well was perforated near the top of the buildup (Fig. 7) to stay away from the oil-water contact and had to be acid stimulated to improve inflow volumes and pressures.

The second well at 9-25 was perforated 14,000 kPa underbalanced using tubing conveyed guns in an attempt to improve formation penetration and clean up perforation damage. Perforations were made near the top and middle of the reef (Fig. 8) and the well required acid stimulation to achieve the desired inflow volumes and pressures.

When the third well, 15-19, was drilled, it was recognized that the high vertical permeability could lead to possible water production from perforations made near the top of the buildup. To lessen the chance of premature water break-through, the 15-19 well was perforated in the middle and lower part of the reef with the lower perforations positioned to stabilize any water coning low in the structure (Fig. 9). Tubing conveyed perforating was used and this time the intervals were perforated totally underbalanced. Using this procedure, flow rates and flowing tubing head pressures similar to those in 14-25 and 9-25 were achieved without acid stimulation. In 15-19, this technique has achieved stabilized water production at the bottom of the producing zone. In addition, the lower perforations should delay the onset of high gas:oil ratios normally seen in perforations made near the top of the reef. Underbalanced perforating not only achieved the production rates desired without acid stimulation, it also eliminated the chance of acidizing into the underlying water [illegible]

At the fourth and final completion at 7-19 the same procedure as 15-19 was used. The middle and lower part of the buildup were perforated totally underbalanced and the desired production rates and pressures were achieved without acid stimulation (Fig. 10).

HYDROCARBON PRODUCTION AND PRODUCTION PROBLEMS

Oil production from the four wells at Hitchcock varies from 18 to 60 m^3/day (Fig. 18). The first well, 14-25, to go on production in October 1989, had, to January 1, 1991, produced 9,860 m^3 (approximately 62,000 barrels) of oil. The most prolific well to January 1, 1991 was 9-25 which in one year produced over 15,000 m^3 (94,500 barrels) of oil (Fig. 18). To January 1, 1991, after only 7 to 15 months of production, the four wells at Hitchcock had flowed 40,120 m^3 (252,346 barrels) of oil.

Production from the Winnipegosis is not without problems. In this and other fields in Saskatchewan, Montana and North Dakota, plugging of perforations and tubing by both salt and paraffin wax is a common problem. In most cases, this is caused by a drop in temperature and pressure at the well bore. Scale, consisting of calcium carbonate, calcium sulphate and zinc sulphate, is also a problem where large volumes of water are produced. Like salt, this precipitates in perforations and tubing and ultimately lowers the productivity of the well. Many remedies, such as injection of hot water, hot oil, scale inhibitors, solvents and acid, have been and are being used in this and other areas.

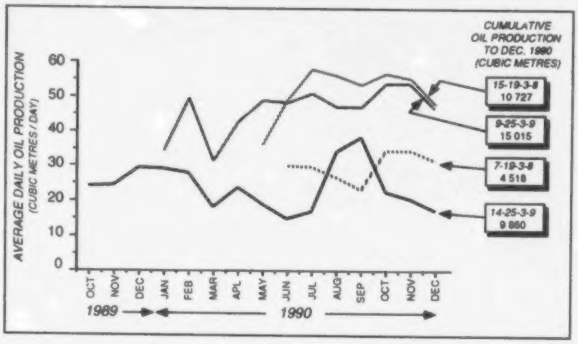


Figure 18: Oil production data for the four Hitchcock wells.

GEOCHEMISTRY OF OILS AND SOURCE ROCKS FROM THE HITCHCOCK AREA

Bituminous mudstones at the base of the Ratner Member in basinal areas between buildups have long been recognized as potentially rich source rocks (Wardlaw and Reinson, 1971). Stoakes *et al.* (1987) and Rosenthal (1987) recognized that the bituminous mudstones, which Stoakes *et al.* (1987) referred to informally as the Brightholme member, are starved basin deposits correlative with Upper Winnipegosis buildups. Although thin (1-3 m) in the Tableland and Hitchcock areas (Fig. 11), Rosenthal (1987) found this unit to be approximately 15 m thick in the Manitoba outcrop belt and Osadetz *et al.* (1990) suggested that a similar thickness could be present in the Estevan area in deeper, undrilled basinal areas between buildups. Brightholme member and equivalent rocks were sampled from three wells in the Hitchcock - Tableland area. Rock - Eval/T.O.C. analyses indicate poor to excellent source rock potential (Table 1).

Table 1: Geochemical data from source rock and crude oil samples from Hitchcock and neighbouring areas.

SOURCE ROCK SAMPLES	DEPTH (m)	T. MAX (°C)	S1(mg/g)	S2 (mg/g)	TR (S1/S1+S2)	T.O.C. (wt. %)	HI	OI
12-24-01-10W2M	2766	445	1.48	2.35	0.38	1.2	195	23
15-07-03-08W2M	2566-81	438	0.04	0.12	0.25	0.13	92	62
07-19-03-08W2M	2546	442	1.32	14.5	0.08	4.3	338	19

OIL SAMPLES	DEPTH (m)	A.P.I. °	% PN	% AROM	PN / AROM	% HC.	Pr / Ph	TS / TM
07-19-03-08W2M	2520	44	74.6	15.2	4.68	90.50	1.02	1.42
15-19-03-08W2M	2500	44	68.7	14.2	4.85	82.9	1.03	1.38
09-25-03-09W2M	2465	43	67.1	16.4	4.10	83.5	1.02	1.18
14-25-03-09W2M	2490	44	61.2	13.8	4.44	74.9	1.04	1.33

KEY TO ABBREVIATIONS

T MAX	Temperature index	A.P.I.	Oil Gravity
S_1	Free hydrocarbons	PN	Paraffin Naphthene
S_2	Residual hydrocarbon potential	AROM	Aromatics
TR	Transformation index	HC	Hydrocarbons
T.O.C.	Total Organic Carbon content	Pr / Ph	Pristane : Phytane ratio
HI	Hydrogen Index	TS / TM	Ratio of 18 alpha (H) / 17 alpha (H) trisnorhopane
OI	Oxygen Index		

Osadetz *et al.* (1990) reappraised the composition of oils analysed by Brooks *et al.* (1987, 1988) from the Williston Basin. They found that a number of oils within a stratigraphic interval from the Silurian Interlake to the Mississippian Ratcliffe have similar molecular compositions. However, the critical characteristics of low C_{21} tricyclic/C_{30} pentacyclic terpanes and a C_{34} hopane predominance serve to distinguish these oils from the two other oil families which occur predominantly in Bakken and younger reservoirs. Abundant pristane, phytane and higher carbon number ($>nC_{20}$) n-alkanes allow these oils to be distinguished from Ordovician oils of similar terpane composition. Using solvent extract molecular compositions, these Winnipegosis oils are all correlatable with Brightholme member source rocks. In fact, all Winnipegosis reef oils belong to this group, termed Family D oils (Brooks *et al.*, 1987; Osadetz *et al.*, 1990).

Values for A.P.I. gravity, gross composition, gasoline range composition, saturate fraction gas chromatographs and terpane mass chromatographs of oils from each of the four Hitchcock wells are similar (Table 1, Fig. 19) and lie within the range for Family D oils. They are, therefore, most likely sourced from the Brightholme member. Like several Family D pools in the Saskatchewan and Madison Group reservoirs, pristane:phytane ratios exceed one. This is the first case where this has been observed in Winnipegosis reservoirs. Significant water washing is inferred from gasoline range hydrocarbon compositions.

Hitchcock oils have the highest A.P.I. gravities found to date in Winnipegosis buildups. Osadetz *et al.* (this volume) have proposed a relationship between the thermal maturity of Winnipegosis source rocks and the gravity of the resulting oils (Fig. 20). In general, there is a linear relationship between oil gravity and heat flow, as a function of depth. For the Hitchcock - Tableland area, heat flow varies from 65 to 70 mW/m^2. Oils that are locally sourced from rocks in the same stratigraphic interval will plot within this envelope (Fig. 20). Any lateral migration of the oil from the source rock will appear as a vertical separation on this graph. The amount of vertical separation provides a measure of lateral migration, probably through the porous Ratner laminites. Oils from Winnipegosis buildups at Bienfait (12-31-3-6W2M) and North Tableland (11-30-2-9 and 4-36-2-10W2M) plot within the source rock evolution envelope and can be inferred to be locally sourced. In contrast, oils from Hitchcock and several Macoun wells have much higher gravities, despite a similar or shallower depth of burial. Their displacement from the source rock evolution envelope is so large that it cannot be reasonably attributed to lateral heat flow variation. Rather, it indicates significant lateral migration from Winnipegosis source rocks deeper in the basin. However, details of the extent and nature of the migration pathway require further investigation.

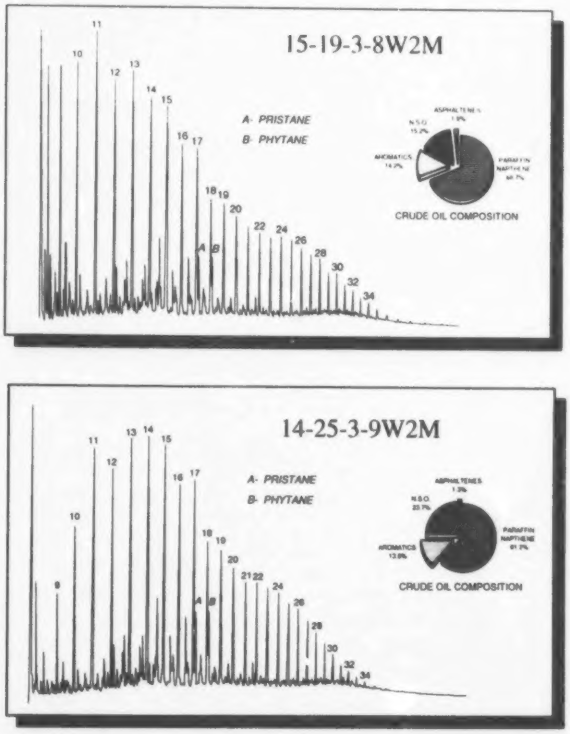


Figure 19: Gas chromatograms and oil analyses of selected Winnipegosis crude oils from Hitchcock.

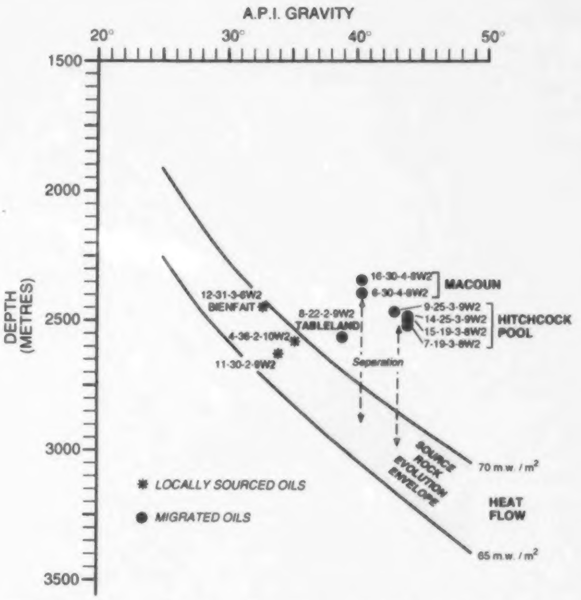


Figure 20: Relationship between oil gravity and burial depth for Hitchcock and neighbouring areas.

CONCLUSIONS

Winnipegosis buildups of the Hitchcock area occur as isolated buildups encased in anhydrite and halite. Large acoustic impedance contrasts at the halite-anhydrite boundary allow the underlying buildups to be mapped seismically. Although conventional 2-D seismic is used to locate the buildups, 3-D seismic is essential to define areal configuration and optimal drilling locations.

Buildups were initiated as a series of codiacean algal shoals which were stabilized by the precipitation of marine fibrous cements. Vertical accretion resulted in entry to the photic zone where corals, bryozoa, red algae and cyanobacteria formed a wave resistant structure with a well-defined lateral facies zonation. Sea level fall and desiccation of the Elk Point Basin caused subaerial exposure of the buildups and formation of a capping calcrete crust. Solution fissures and cavities formed at this time were filled with vadose silt.

Dolomitization occurred by seepage reflux of magnesium-rich marine waters. Later recrystallisation and dissolution of this early dolomite enhanced porosity locally. Reservoir quality of the Winnipegosis varies throughout the buildup and is related to both primary depositional facies and subsequent diagenetic events. In general, reef interior facies exhibit the best reservoir quality. Porosity and permeability are reduced by anhydrite in the laminated dolostone facies, by early marine fibrous cements in the shoal margin subfacies and by vadose silt in the upper part of the buildup wherever primary and secondary pores, cavities and fissures have been infilled.

Geochemical analysis of oils and source rocks from Hitchcock and neighbouring areas has shown that the informally named

Brightholme member at the base of the inter-reef laminites is the source of all Winnipegosis oil in this area. Not all of the buildups are locally sourced, however, and studies have shown that the oil at Hitchcock has migrated some distance and is slightly degraded. Winnipegosis buildups are capable of significant hydrocarbon production given a suitable source rock and migration pathway. However, production problems such as salt, wax and scale precipitation and the coning of formation water can contribute to both short and long term loss of production.

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PETROLOGY OF THE MIDDLE BAKKEN MEMBER IN THE DALY FIELD, SOUTHWESTERN MANITOBA

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ABSTRACT

The Bakken Formation is a thin but very widespread Mississippian-Devonian siliciclastic unit. In much of the Williston Basin three members can be recognized: a relatively coarse sandstone/siltstone unit (Middle Member) separates two black to brown, organic-rich shales (Upper and Lower Members). Because of their characteristic high radioactivity, the Upper and Lower Members are excellent subsurface markers. The Middle Bakken is an oil reservoir in North Dakota, Saskatchewan, and Manitoba.

In the Daly field in southwestern Manitoba, the Lower Bakken is absent and the Middle Member shows a sharp erosional contact with the underlying Devonian Lyleton Formation. The Middle Bakken Member consists of four lithofacies: i) wavy laminated-lenticular bedded sandstone; ii) horizontally laminated siltstone; iii) massive, bioturbated siltstone; and iv) breccia. These rocks are composed mainly of quartz, clay minerals, and dolomite, with minor amounts of anhydrite, calcite, halite, feldspars, hematite, and pyrite. The amount of matrix material (mainly clay minerals) generally increases upwards in the section, with a concomitant decrease in quartz content. Similarly, the proportion of authigenic dolomite increases from less than 20 percent at the base to over 40 percent at the top of the member.

Thin section and SEM examination show that the Middle Bakken Member underwent at least five diagenetic stages. The first stage involved dolomite formation primarily as small, euhedral rhombohedra coating detrital grains and filling pores. This occurred before substantial porosity reduction by compaction and most likely took place either penecontemporaneously with sedimentation or at shallow depths below the sediment-water interface during very early methanogenesis. At the same time, pyrite was precipitated locally and in some cases forms marker horizons that can be traced from well to well. During the second stage, illite was precipitated as rims and coats around the detrital grains. Minor secondary intragranular porosity formed as a result of dissolution of the authigenic dolomite crystals and detrital feldspar grains. The third stage is marked by the formation of quartz overgrowths on detrital grains. Following this, neoformed K-feldspar was precipitated. The final stage was the formation of anhydrite cement. These last two episodes were likely associated with the oxidation of pyrite to form elevated sulphate/low pH conditions, and the subsequent reaction of illite to form K-feldspar and the precipitation of calcium sulphate.

INTRODUCTION

The Bakken Formation, a thin, widespread siliciclastic unit of Mississippian-Devonian age, is made up of three members across much of the the Williston Basin. At the base, black to brown, organic-rich shales (Lower Member) are overlain by, in turn, a sandstone/siltstone unit (Middle Member) and dark-coloured, organic-rich shales (Upper Member). Both shaly members are useful marker horizons, easily recognizable on gamma-ray logs by their characteristically high radioactivity. The Middle Bakken is an oil reservoir in North Dakota, Saskatchewan, and Manitoba.

In Manitoba, the Lower Bakken is absent except for a small area in the southernmost part of the province (McCabe, 1959, 1971; Martiniuk, 1988). Elsewhere, the Middle Bakken sandstones and siltstones unconformably overlie dolomitic green and red shale of the Devonian Lyleton Formation (Three Forks in North Dakota and Montana, and Torquay in Saskatchewan). The Upper Bakken is overlain by the Lodgepole Formation, a cherty argillaceous limestone. The contact between the Lodgepole and the Bakken appears to be conformable in Manitoba, but Kent (1984) has identified an unconformity between the two formations in Saskatchewan. The edge of the Bakken Formation is truncated by the Mississippian unconformity.

The Bakken Formation is generally considered to be equivalent, in terms of both chronostratigraphy and depositional setting, to other widespread Devonian-Mississippian black shale units of North America (*e.g.*, Exshaw Formation of the Alberta Basin; Chattanooga Shale of the mid-continent region; Antrim Formation of the Michigan Basin; Meissner, 1978; Thrasher, 1987). Indeed, the formation, overall, fits well with the general stratigraphic/depositional sequence for the Devonian-Mississippian boundary (Gutschick and Moreman, 1967; Richards, 1989) which comprises Devonian carbonates (Lyleton-Three Forks) overlain by Late Devonian-Early Mississippian black shale (Lower Bakken), followed by Early Mississippian sandstones, siltstones, (Middle Bakken), and shales (Upper Bakken), and ending with Mississippian carbonates (Lodgepole).

One of the most definitive examinations of the Bakken Formation in Manitoba is that of McCabe (1959). More recently, Martiniuk (1988) provides a comprehensive summary of the subsurface geology and petroleum potential of the formation, incorporating all of the subsurface data (over 400 wells) available through 1987. Arbez (1988) presents the results of a detailed petrophysical study of the Middle Bakken reservoir in the Daly field. The purpose of this paper is to summarize the preliminary results of a detailed sedimentological and petrographic study of the Bakken sequence in the Daly field area of southwestern Manitoba.

STUDY AREA AND METHODOLOGY

The Daly field is located in southwestern Manitoba, immediately adjacent to the Saskatchewan border between Townships 8 and 10 (Fig. 1). All available cores of the Bakken Formation within the field and surrounding area were examined to determine lithofacies, depositional setting, and diagenesis. A total of 122 core samples from 34 wells was analyzed for bulk mineralogy, detailed carbonate mineralogy, clay mineralogy, texture, and fabric using standard X-ray diffraction, thin section, and scanning electron microscopy-energy dispersive X-ray techniques. Grain size distribution and textural parameters of the sandstones were also estimated using an automated particle size analyzer (Brinkman PSA 2010) coupled with a Galai image analysis system. Edwards (1991) and Edwards and Last (1989) have provided details of analytical procedures, preparation and quantification techniques, and instrumentation.

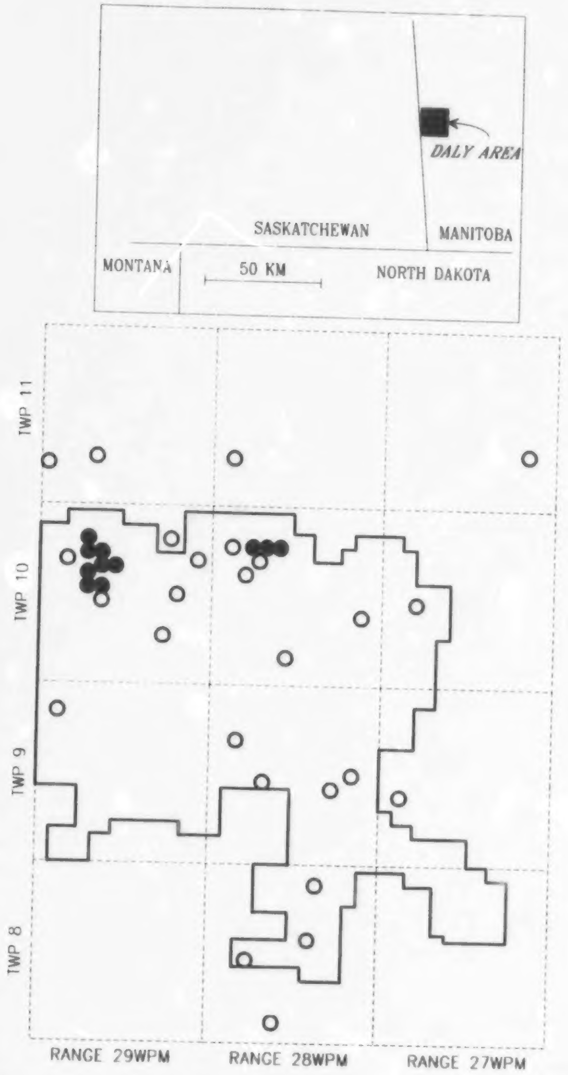

Figure 1: Map showing the location of the Daly field area in southwestern Manitoba and the cores of the Bakken Formation sampled in this study. Open circles are cores from nonproductive wells; closed circles are cores from producing Bakken oil wells.

LITHOFACIES

The Bakken Formation in Manitoba can be described in terms of five major lithofacies on the basis of lithology, sedimentary structures, texture, colour, and fossil content. The contacts between the lithofacies are generally gradational, although the lower contact of Lithofacies 5 (Breccia) is always sharp. Dolostone and limestone lithotypes which are minor components of the Bakken section in the Daly area are not included here.

LITHOFACIES 1: SHALE AND MUDSTONE

This lithofacies consists of organic-rich, fissile to platy, laminated shale and structureless mudstone. Colours are black (N1) to brownish black (5YR2/1) and dark dusky red (10R2/2). The rock often breaks with a smooth, conchoidal fracture. The core commonly breaks along bedding planes into "poker-chip"-shaped pieces. When laminae are present, the sediments typically exhibit even, parallel, very thin (less than 0.5 mm) lamination. Both calcareous and pyritic thin beds/laminae are present.

LITHOFACIES 2: MASSIVE SILTSTONE

The rocks of Lithofacies 2 comprise homogeneous, olive grey (5Y4/1) sandy siltstone and silty sandstone with abundant fossils. In some cores, this lithofacies exhibits a dusky red colour (5R3/4), and a mottling of red (5R5/4), green (5GY5/2), and grey (5Y4/1) colours, presumably due to postdepositional oxidation. The macroscopic fossil remains are primarily brachiopods with some crinoid stem fragments. The sediment appears bioturbated and churned; however, distinct burrow traces are rare. Although usually well lithified, in some cores this facies has a weakly indurated, crumbly character.

LITHOFACIES 3: HORIZONTALLY LAMINATED SILTSTONE

The sandy siltstones and silty sandstones of this lithofacies are characterized by well-developed, thin to medium (0.5 mm to 2 mm), very low-angle to horizontal parallel laminations. The sediments are less commonly very thin bedded (10 mm to 2 cm) and show indistinct ripple crossbedding. In some cores, the lithofacies exhibits a fining upward sequence, grading from very thin bedded, crossbedded, coarse siltstone and fine sandstone at the base to thin, parallel horizontally laminated, very fine sandstone and siltstone at the top. Colours vary from olive grey (5Y4/1) to dusky red (5R3/4) and light brown (5YR5/6).

LITHOFACIES 4: WAVY LAMINATED-LENTICULAR BEDDED SANDSTONE

The sediments of this lithofacies comprise coarse silty sandstones and fine to medium grained sandstones characterized by wavy to lenticular bedding and, more rarely, well-developed, small scale ripple crossbedding. Sediments are a uniform medium grey (N5) to olive grey (5Y4/1) except when oil stained. This facies exhibits good (10 to 20 percent) intergranular porosity in some of the cores.

LITHOFACIES 5: BRECCIA

This lithofacies is a variable assemblage of carbonate, shale, and sandstone pebbles in a silty to very fine grained sandstone matrix. In several cases the matrix material is dolomitic mudstone or argillaceous dolostone. The pebbles are generally angular to subangular and often show a preferred orientation. The rocks are dominantly matrix supported and show varying degrees of sorting. The facies is characterized by a sharp (scoured?) basal contact.

LITHOFACIES SEQUENCE

In most of the Daly area, the Bakken sequence can be described as a single succession of lithofacies. Lithofacies 5 only occurs in the lowermost part of the Middle Bakken Member, marking the unconformable contact with the underlying Lyleton Formation. The remainder of the Middle Bakken grades from the relatively coarse clastics of Lithofacies 4 upward through well-laminated siltstones of Lithofacies 3 into an interval dominated by intercalation of Lithofacies 2 and 3. The Upper Bakken is composed of the shales and mudstones of Lithofacies 1. With the exception of the sharp (erosional?) contact at the base of the section, no evident unconformities are recognized in this Bakken sequence in the Daly area.

INTERPRETATION OF LITHOFACIES

Much of the previous sedimentological work on the Bakken has concentrated on the Upper and Lower Members of the formation. Most investigators suggest that these shales were deposited in a relatively shallow water setting (Martiniuk, 1988; Kent, 1984; McCabe, 1959; Raasch, 1956), with possible environments ranging from a vast marine swamp with restricted circulation to a marginal marine lagoon to a broad open marine shelf/platform over which wave intensity was dampened due to shallow depths. In contrast, others have presented convincing arguments for deep water, open marine, sediment-starved basin deposition (Lineback *et al.*, 1987; Webster, 1984; MacDonald, 1956).

Within the Daly area, the grain size, lamination, and high organic content of Lithofacies 1 all indicate deposition in low energy (below wave base), anaerobic conditions. The poorly laminated to massive, bioturbated and churned siltstones of Lithofacies 2 were most likely deposited under slightly shallower, dysaerobic conditions, but still below effective normal and storm wave base, whereas the well-laminated and crossbedded siltstones of Lithofacies 3 record deposition in a setting affected at least periodically by wave energy. Interbedding of Lithofacies 2 and 3 throughout much of the Middle Bakken section reflects slight fluctuations in wave base, possibly due to vertical changes in stratification of the overlying water column. The relatively coarse grain size and well bedded nature of Lithofacies 4 indicate deposition in a moderate energy environment characterized by variable current velocity. The breccia of Lithofacies 5 was deposited in a setting that was subject to high erosive current action and very short transport distances.

DEPOSITIONAL HISTORY

The first record of Bakken sedimentation in the Daly area of Manitoba is the coarse- to fine-grained sandstones and siltstones of the Middle Member of the formation. Black, organic-rich shales of the Lower Bakken were probably deposited in the area, presumably under quiet water, relatively deep basin, marine anaerobic conditions, but subsequent regression of the seas associated with a regional eustatic sea level fall resulted in erosion of these sediments from nearly all of the province. Renewed transgression resulted in deposition of a generalized fining upward sequence in the Middle and Upper Bakken units from the wavy and lenticular bedded sandstones of Lithofacies 4, through horizontally laminated and bioturbated siltstones of Lithofacies 3 and 2, and, finally, into black, organic-rich shales and mudstones of Lithofacies 1.

PETROGRAPHY

The Bakken Formation is characterized by a wide range of lithologies and textures. The "average" sandstone of the Middle Bakken Member (Lithofacies 2 to 4) is a dolomitic arkosic wacke (terminology after Dott, 1964, as modified by Pettijohn, 1975). However, a considerable range of sandstone types is present, including rocks of quartz arenite and quartz wacke composition. The bulk mineralogical composition of Lithofacies 1 to 4 is summarized in figure 2.

FRAMEWORK GRAINS

Quartz normally makes up about 70 percent of the framework grains in the sandstones. Nearly all of this quartz is monocrystalline and occurs mainly as subrounded to subangular grains. The amount of quartz generally decreases upward in the section (Fig. 3). About 13 percent of the mineral grains are feldspar, dominantly orthoclase. Most of the feldspar grains in thin section and hand samples are subrounded and show rims of chemical alteration. Scanning electron microscope examination also reveals that the feldspar grains commonly show considerable surface corrosion and dissolution pitting (Fig. 4).

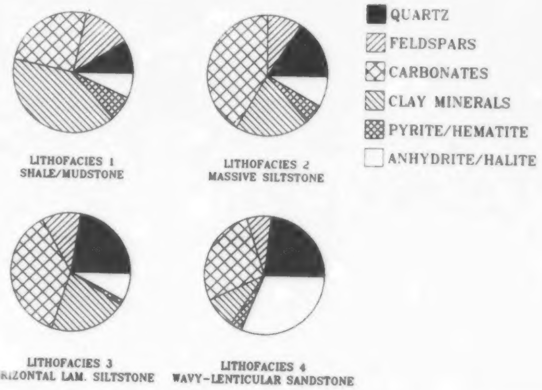

Figure 2: Pie diagrams showing the bulk mineralogical composition of the Bakken Formation lithofacies identified in this study.

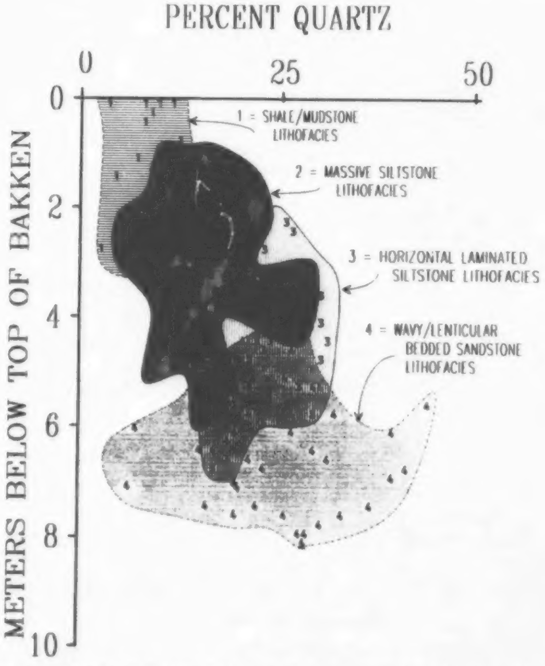

Figure 3: Plot showing the change in amount of quartz versus depth in the Bakken Formation.

MATRIX

Except for some of the samples from Lithofacies 4, nearly all Bakken sandstones have a high matrix content, with an overall average of nearly 35 percent of the total rock. Matrix, as discussed here, includes all materials which are too fine to be readily identified in thin section petrography. Thus, it may include material that is either detrital or authigenic in origin. The matrix most commonly observed in the Bakken sediments is an indistinct mixture of fine silt-sized quartz, mica, and feldspars, detrital clays, and authigenic clay and carbonate minerals. In general, the proportion of matrix material in the rocks increases upward in the Bakken section. The sandstones from Lithofacies 4 have much less matrix material than other Middle Bakken facies.

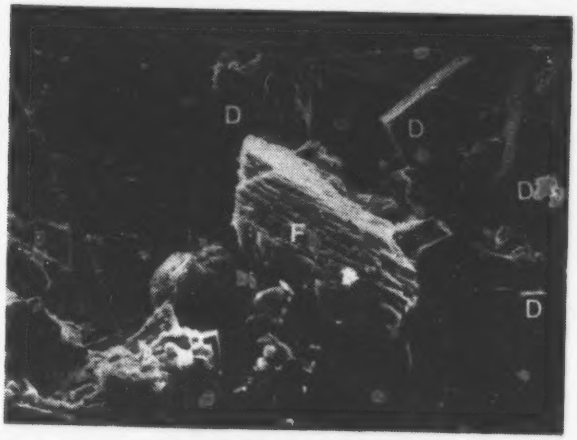

Figure 4: Scanning electron photomicrograph showing partial dissolution and etching of a detrital feldspar grain (F) from Lithofacies 4. Note also the euhedral and partly dissolved dolomite rhombs (D).

AUTHIGENIC MINERALS

Dolomite and anhydrite are the most important cementing agents in the Bakken Formation sandstones, generally forming about nearly 30 percent of the rock volume. Other authigenic minerals include: quartz, clay minerals, feldspars, pyrite, hematite, and calcite.

Dolomite

Dolomite cement is a major component of the rocks of the Bakken Formation and occurs predominantly as small (2 to 20 μm), interlocking, euhedral rhombs which coat detrital grains and infill intergranular pores. Larger (30 to 50 μm) rhombs are also present. These large rhombs, which are often zoned, occur as interlocking aggregates in pores and as displacive crystals in matrix material. Although most of these large dolomite crystals are euhedral, corrosion and dissolution pitting on the surfaces of some of the rhombs indicate the material was, at some point, in contact with relatively acidic pore fluids (Fig. 5).

The finely crystalline dolomite appears to pre-date all other authigenic minerals in most samples. The larger dolomite rhombs, however, formed later in the paragenetic sequence, subsequent to mechanical compaction, clay mineral and feldspar authigenesis, and precipitation of iron oxides and sulphides.

Lithofacies 2 and 3 have, overall, the highest proportion of dolomite cement (Fig. 2). Although the sandstones are generally well-cemented by dolomite, these rocks often contain abundant micro-intercrystalline porosity.

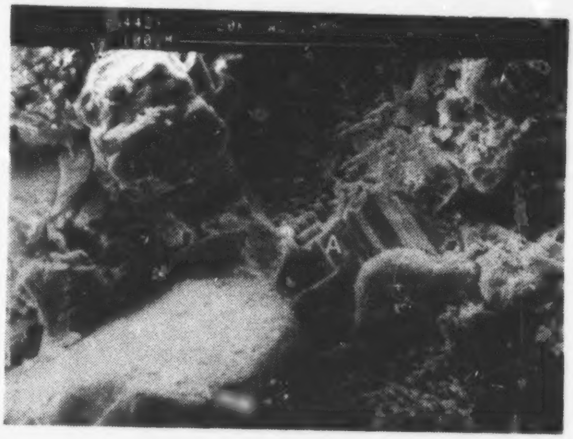

Figure 5: Scanning electron photomicrograph showing partial dissolution of an authigenic dolomite rhomb from Lithofacies 3.

Calcite

Authigenic calcite is present only rarely in the Middle Bakken sediments. It occurs as small, euhedral crystals associated with the coarse dolomite and appears to have formed at the same time or immediately after the dolomite.

Pyrite and Hematite

Iron sulphide and iron oxide minerals occur widely in the Bakken rocks as very small (less than 2 μm), euhedral octahedra and pyritohedra, and larger (10 to 15 μm) aggregates and framboids. Morphologically, the two minerals are indistinguishable under SEM and are differentiated only on the basis of the presence or absence of sulphur in the EDS spectra. In some samples, authigenic pyrite and hematite show obvious transformation and replacement relationships with pre-existing detrital grains and fossils. Pyrite also occurs as a cement in a laterally extensive bed in Facies 2. This 10 cm thick bed can be readily identified on lithodensity logs in parts of the Daly area. Most of the pyrite formed either syndepositionally at the sediment-water interface under strongly anoxic conditions or possibly very early in the postdepositional history of the rock. The formation of hematite, which required oxidizing conditions, may be due to alteration of detrital ferromagnesian silicates or early-formed diagenetic pyrite. Clear evidence to support either an early or late-stage genesis of hematite is lacking.

Clay Minerals

Illite (Fig. 6) is the most abundant authigenic clay mineral and is readily identified in both thin section and SEM examination of the Bakken sediments. It generally occurs as irregular flakes and delicate lath-like and fibrous spines. The illite commonly forms coats and rims around detrital grains but post-dates most of the dolomite and iron sulphide formation. Small amounts of authigenic kaolinite have also been identified in the sediments. This kaolinite appears as very small (less than 2 μm) platy parallel intergrowths associated with pyrite crystals and framboids and most likely formed at the same time as the pyrite.

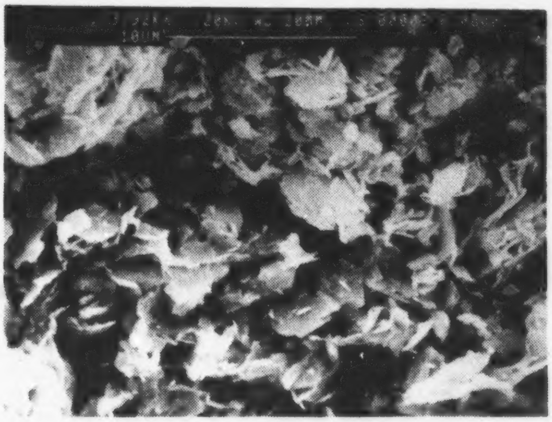

Figure 6: Scanning electron photomicrograph showing flaky authigenic illite from Lithofacies 2.

Quartz

Authigenic quartz overgrowths occur in many of the Bakken sandstone samples but are volumetrically negligible. Much of the quartz precipitation occurred subsequent to the initial phase of dolomite formation. Some samples, however, show that partial interlocking of quartz overgrowths took place prior to carbonate precipitation.

K-Feldspar

Authigenic feldspars are mostly K-feldspars occurring both as euhedral, neoformed crystals precipitated directly from pore fluids and as replacement and alteration products of pre-existing minerals. Aggregates of finely crystalline, euhedral K-feldspar occur in minor amounts in most sandstones. The fabric relationships of these crystals indicate that they are relatively late-stage precipitates, post-dating the formation of quartz, pyrite, and carbonate minerals.

Anhydrite

Anyhdrite commonly occurs as large, lath-shaped poikilotopic pore-filling crystals (Fig. 7). This cement lithifies both clay-coated and dolomite-rimmed detrital grains, earlier-formed iron oxides, corroded detrital feldspar grains and euhedral K-feldspars. Therefore, precipitation of anhydrite is the latest diagenetic process in the post-depositional sequence of the formation.

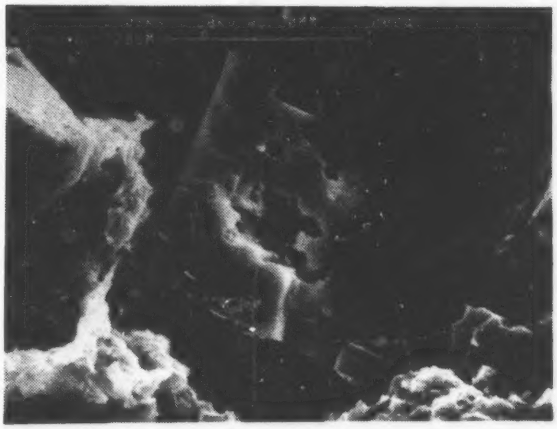

Figure 7: Scanning electron photomicrograph showing coarsely crystalline pore-filling anhydrite (A) from Lithofacies 4.

Anhydrite cements are most common in the sandstones of Lithofacies 4 (Fig. 2) and in the breccia of Lithofacies 5.

DIAGENETIC HISTORY

The diagenetic history of the Bakken sediments began with the precipitation of very early, finely crystalline dolomite and iron sulphide cements. The formation of these authigenic minerals was very likely enhanced by microbial reduction of sulphate ions in the pore waters under anaerobic conditions (*i.e.*, $SO_4^{2-} + 2CH_2O + H_2O \rightarrow H_2S + 2HCO_3^-$). As the chemistry of the pore fluids changed from highly alkaline (high pH) to more acidic (low pH) solutions with increased burial, some dissolution and corrosion of detrital feldspars and authigenic carbonates occurred. This partial degradation of detrital feldspar grains may have provided a possible source for potassium, alumina, and silica which were precipitated as authigenic illite and, in minor amounts, as quartz overgrowths. Subsequent oxidation of pyrite, possibly by downward percolating groundwaters associated with the development of a post-Mississippian unconformity, resulted in reaction of some of the authigenic illite to form K-feldspar. The oxidizing conditions also resulted in the formation of hematite and the distinctive red coloration and mottling found in the Middle and Upper Bakken sediments. Finally, the formation of coarsely crystalline calcium sulphate cement marked the last major phase in the diagenesis of the Bakken. This late-stage anhydrite is facies selective, forming in parts of the unit that had the highest remaining primary intergranular porosity and permeability.

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CONODONTS OF THE BAKKEN FORMATION (DEVONIAN - MISSISSIPPIAN) IN SASKATCHEWAN, NORTHERN WILLISTON BASIN

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ABSTRACT

The Bakken Formation comprises a grey sandstone bed between two distinctive, radioactive black shales. It straddles the Devonian-Mississippian boundary and is recognized as a premier stratigraphic marker and hydrocarbon source rock in the Williston Basin. Conodonts were observed scattered on bedding planes in both the upper and lower Bakken shales from 11 wells. Thirteen samples containing conodonts were collected in an effort to determine the age and thermal maturity of the formation. The conodonts are placed in 21 taxa, consisting mostly of platform elements of *Siphonodella*, *Polygnathus*, *Branmehla*, *Bispathodus*, *Pseudopolygnathus*, *Palmatolepis* and *"Spathagnathodus"*. Conodont evidence indicates that the Bakken is as old as the Late Famennian Lower *Palmatolepis gracilis expansa* Biozone and as young as the Kinderhookian Lower *Siphonodella crenulata* Biozone. The Bakken sandstone lacks precise age-diagnostic conodonts, but paleobotanical evidence indicates that the Devonian-Mississippian boundary may coincide with the base of unit 3 of the Bakken sandstone in Saskatchewan. Conodont and lithologic evidence suggest that the two Bakken shales were deposited in a euxinic, reducing, offshore marine environment under conditions of very slow rates of sedimentation.

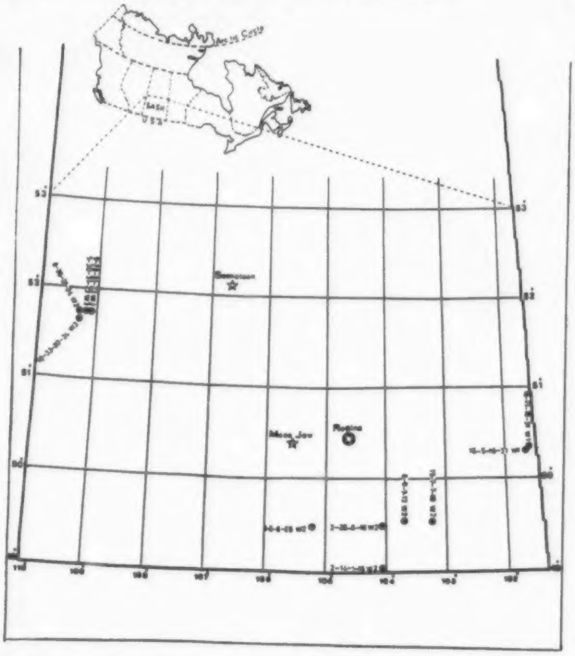

Figure 1: Conodont sample location map.

INTRODUCTION

The Bakken Formation was initially considered Mississippian in age (Nordquist, 1953) but paleontologic evidence has indicated that the Bakken is of both Devonian and Mississippian age (*i.e.* it is transitional) (Christopher, 1961; Holland *et al.*, 1987; Kent, 1987; Thrasher, 1987). Prior to this study conodont evidence for the age of the Bakken in Saskatchewan has come from sources outside the province (Hayes, 1985). Thirteen samples have been collected from both the upper and lower Bakken shales in 11 borehole cores (Fig. 2). None was recovered from the Bakken middle sandstone. Initial examination was done with aid of a hand lens (10x) and binocular microscope and a more detailed identification was carried out by Dr. Tom Uyeno (1989) of the Geological Survey of Canada, Calgary, whose help is herewith gratefully acknowledged. The samples are stored at the Institute of Sedimentary and Petroleum Geology in Calgary.

CONODONT BIOZONATION

The standard Upper Devonian-Lower Mississippian conodont zonation (Fig. 3) was adapted to an interregional stratigraphic analysis based on depositional complexes by Sandberg (1979) and Ziegler and Sandberg (1984). The base of each zone is usually defined by the stratigraphic first occurrence of a particular taxon. Phyletic changes within a single genus *Palmatolepis* are the basis for most of the Upper Devonian conodont zones, whereas the Kinderhookian zonation is based on species of *Siphonodella*. Zeigler and Sandberg (1984) concluded that the first occurrence of *Siphonodella sulcata* should be used to mark the Devonian-Mississippian boundary. This zone is dominated by *Bispathodus aculeatus aculeatus*, *Bispathodus stabilis* and *Polygnathus communis communis*.

COMPOSITION AND DISTRIBUTION OF THE CONODONTS

The Bakken conodonts of this study are placed in 21 taxa. The presence and distribution of taxa, sample depth and location are shown in Figure 3. The collection is composed mostly of platform (pectiniform) elements and rare ramiform (barlike) elements. The platform genera include *Siphonodella*, *Polygnathus*, *Brenmehla*, *Bispathodus*, *Pseudopolygnathus*, *Palmatolepis*, and *"Spathognathodus"*. All conodont specimens were observed as lags which are concentrated parallel to bedding surfaces and which contain fish bones and teeth, wood fragments, and phosphatic particles. Two thirds of the collection were obtained from the upper shale member, the remainder from the lower shale member.

CONODONT TAXA	LOWER SHALE	UPPER SHALE	DEPTH (m.)	SAMPLES & WELL LOCATION
Bispathodus aculeatus aculeatus		●	1675.3	6-9-7-13W2
		●	2324.7	2-14-1-16W2
		●	792.75	9-29-31-23W3
		●	844-845	10-33-30-24W3
Bispathodus jugosus	●		1819.3	1-2-6-25W2
	●		849.7-851	10-33-30-24W3
Bispathodus stabilis		●	792.75	9-29-31-23W3
		●	844-845	10-33-30-24W3
Branmehla inornata	●		1689.6	15-7-7-10W2
	●		1819.9	1-2-6-25W2
	●		835.3	5-30-31-23W3
	●		849.7-851	10-33-30-24W3
Palmatolepis g.sigmoidalis	●		1819.9	1-2-6-25W2
	●		849.7-851	10-33-30-24W3
Palmatolepis perlobata postera	●		849.7-851	10-33-30-24W3
Palmatolepis rugosa rugosa	●		849.7-851	10-33-30-24W3
Polygnathus communis communis		●	686.8	16-5-16-31W1
		●	1675.8	6-9-7-13W2
		●	2324.7	2-14-1-16W2
		●	792.75	9-29-31-23W3
		●	844-845	10-33-30-24W3
Polygnathus inornatus		●	1675.3	6-9-7-13W2
Polygnathus ? sp.		●	811.4	8-26-31-24W3
Pseudopolygnathus marginatus		●	654.1	16-11-16-31W1
		●	1675.8	6-9-7-13W2
		●	2324.7	2-14-1-16W2
		●	792.75	9-29-31-23W3
		●	844-845	10-33-30-24W3
Pseudopolygnathus primus		●	1675.3	6-9-7-13W2
Pseudopolygnathus ch.p. brevipennatus	●		849.7-851	10-33-30-24W3
Pseudopolygnathus sp.		●	792.75	9-29-31-23W3
Siphonodella sp.		●	792.75	9-29-31-23W3
Spathognathodus crassidentatus		●	792.75	9-29-31-23W3
Spathognathodus strigosus	●		1819.9	1-2-6-25W2
	●		849.7-851	10-33-30-24W3
Spathognathodus sp.	●		1689.4	15-7-7-10W2
	●		2324.5	2-14-1-16W2
	●		849.7-851	10-33-30-24W3
Spathognathodus elongatus		●	811.4	8-26-31-24W3
Spathognathodus macer		●	1675.3	6-9-7-13W2
		●	2324.7	2-14-1-16W2
		●	792.75	9-29-31-23W3
		●	844-845	10-33-30-24W3
Ramiforms		●	1784.95	3-26-6-16W2
		●	811.4	8-26-31-24W3
	●		2324.5	2-14-1-16W2

Figure 2: Conodont biozonation. Lower Mississippian zonation after Sandberg (1979, Fig. 3) and Upper Devonian zonation shows comparison of "new" (Ziegler and Sandberg, 1984) and "old" (Sandberg, 1979) zonation schemes.

LOWER SHALE

Based on the ranges established elsewhere (Sandberg, 1979; Ziegler and Sandberg, 1984; Holland *et al.*, 1987), the most diagnostic taxa found in the lower Bakken shale are *Palmatolepis gracilis signoidalis*, *Palmatolepis rugosa rugosa* and *Bispathodus jugous*. Sandberg (1979) and Ziegler and Sandberg (1984) reported that the Lower *expansa* zone is defined by the appearance of *Palmatolepis gracilis signoidalis*, *Palmatolepis rugosa rugosa*, *Pseudopolygnathus brevipennatus*, and *Bispathodus jugous*. However, the previously unrecognized *Bispathodus aculeatus* zone and the presence of *Brenmehla inornata* (="*Spathognathodus*" *inornatus*) also indicate that the lower Bakken shale in Saskatchewan can be assigned to the Lower-Middle *expansa* zone. Thus the Lower Bakken shale is Late Famennian.

UPPER SHALE

Since species identification requires a complete discrete specimen, the conodont evidence from this member is not precise because the specimens of *Siphonodella* are fragmented and whole recoveries were rare. Sandberg (1979) reported that the Lower *Siphonodella crenulata* zone can be recognized by *Pseudopolygnathus marginatus*. The recovered upper shale fauna include: *Pseudopolygnathus marginatus*, *Bispathodus aculeatus aculeatus*, "*Spathognathodus*" *macer and Polygnathus inornatus*. These suggest that the upper Bakken shale is no younger than the Lower *Siphonodella crenulata* zone (Middle Kinderhookian). Hayes (1985) and Holland *et al.* (1987) also placed their fauna in the Lower *Siphonodella crenulata* zone.

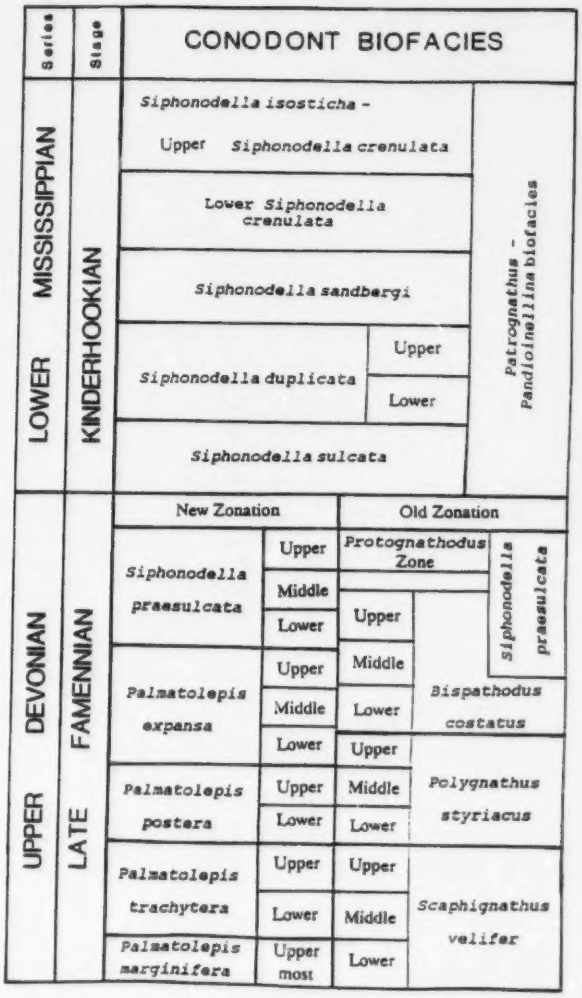

Figure 3: Distribution of conodonts obtained from the lower and upper Bakken shales. Depths are from the core box and well location related to Figure 1.

CONODONT COLOUR ALTERATION INDEX (CAI)

Epstein *et al.* (1977) determined that the alteration in conodont colour from pale yellow to black is both time and temperature dependent and this colour transformation can be used as a geothermometer to evaluate the hydrocarbon generation potential of the conodont host rock. They caution that a colour alteration index for conodonts does not define thermal thresholds of oil generation, but does indicate thermal cutoffs for oil generation and the onset of condensate and dry gas production. Harris (1981) cites that conodont colour alteration begins beyond the threshold of hydrocarbon generation, and a CAI of 1.5 to 2 represents the threshold of oil and condensate production. Conodonts from both the upper and lower Bakken shales have CAI values of 1 (pale yellow) to 1.5 (very pale brown) as noted by Uyeno (1989). Since the conodont samples were collected from depths ranging between 800 and 2,350 m, the CAI values suggest that the host rock is thermally immature to poorly mature between these depths. However, the use of the CAI method is preliminary to evaluation of the Bakken as a hydrocarbon source rock, and hence further investigations of thermal maturity of the Bakken shales are considered desirable.

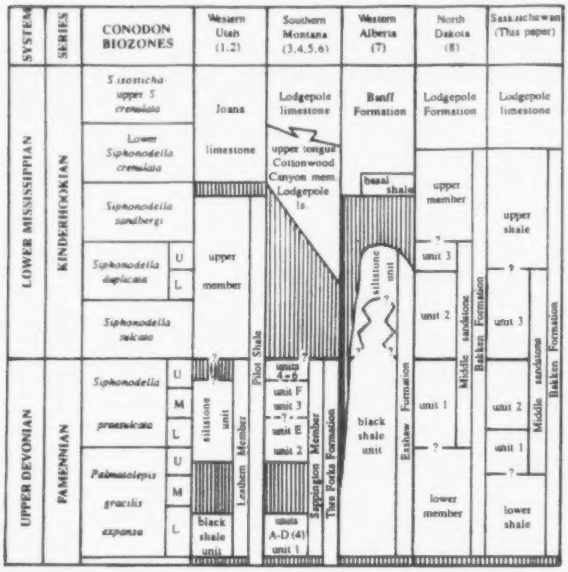

Figure 4: Correlation of the Bakken Formation with its equivalents in the North America Western Interior (modified from Holland *et al.*, 1987). Conodont biozones from Sandberg (1979) and Ziegler and Sandberg (1984). Sources of information for each column correspond to numbers enclosed in parentheses as follows: (1) Gutschick and Rodriquez (1979); (2) Sandberg *et al.* (1980); (3) Sandberg *et al.* (1972); (4) Gutschick *et al.* (1962); (5) Sandberg (1956); (6) Sandberg and Klapper (1967); (7) Macqueen and Sandberg (1970; (8) Holland *et al.* (1987).

DISCUSSION

Conodont evidence from the Bakken Formation in Saskatchewan indicates that portions of the Bakken are correlative with the Exshaw Formation in Alberta, Sappington Member of Three Forks Formation in Montana, Leatham Member of the Pilot Shale in Utah and Nevada (Fig. 4), as well as other equivalent rocks (Cottonwood, Englewood, Leatham) stretching across the North America Western Interior (Holland *et al.*, 1987). Although, the present conodont evidence is insufficient to document the systemic Devonian-Mississippian boundary, due to scarce recovery of conodonts between the Lower *expansa* zone (lower shale) and the Lower *Siphonodella crenulata* zone (upper shale), abundant plant impressions, *i.e.*, blade-like leaves, in unit 3 of the middle sandstone (Karma, 1991) indicate that the appropriate Devonian-Mississippian boundary may coincide with the base of this unit in Saskatchewan. Christopher (1961) placed this boundary at the base of the Bakken sandstone. Hayes (1985) located it close to or at the base of the upper shale. Holland *et al.* (1987) placed the boundary at the base of unit 2 of the middle sandstone in North Dakota.

On the basis of lithologic and paleontologic evidence, Karma (1991) suggested that the Bakken shales were deposited in an euxinic, offshore, marine environment under conditions of very slow rates of sedimentation. This is confirmed by the presence of the *Palmatolepis-Polygnathus* biofacies and the *Siphonodella* and *Pseudopolygnathus* species which were deposited in relatively deep water (Sandberg, 1976; Sandberg and Gutschick, 1983; Hayes, 1984). Bedding features and fossils from the sandstone member indicate a current-influenced, aerobic, shallower water, marine environment.

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PETROLEUM POTENTIAL OF THE MIDDLE MEMBER, BAKKEN FORMATION, WILLISTON BASIN

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ABSTRACT

The Bakken Formation is a widespread clastic unit between the Mississippian Madison Group and the Upper Devonian Big Valley, Torquay, or Three Forks formations. The formation typically consists of three members: a lower and upper organic-rich shale and a middle calcareous sandstone and siltstone.

The three members of the Bakken Formation exhibit an onlapping relationship, and converge and thin toward the margins or marginal shelf areas of the Williston Basin. In Manitoba, the lower member is absent, except for Waskada field.

Several correlative lithofacies have been recognized in the middle member. The lowermost unit of the middle member is a regressive deposit. The upper portions of the middle member

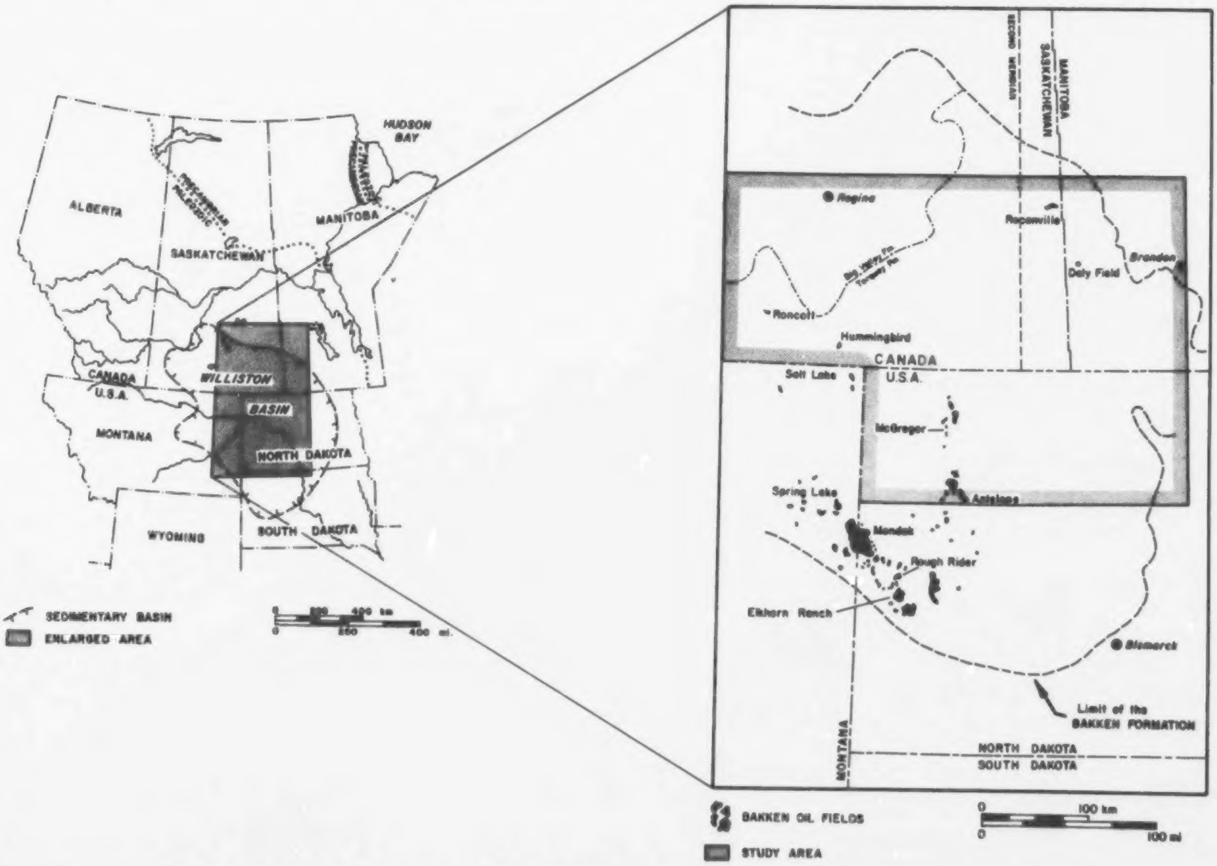


Figure 1: Index map showing the location of the study area within the Williston Basin. The approximate limit of the Bakken, Torquay, and Big Valley formations is indicated. Oil fields which are producing or have produced from the Bakken Formation are also shown.

show transgressive and regressive cycles, and intertidal deposits interfinger with their distal equivalents.

Conventional Bakken production is restricted to areas where it is a thermally mature source rock. The greater part of Bakken production in recent years has been taken from the shale of the upper and lower members, although there are outlying fields which produce from a sandstone within the middle member. This suggests that there has been a migration of Bakken oil from the central portion of the basin. The sandstone unit may provide a conduit for migration.

INTRODUCTION

The Bakken Formation has long been considered to be an important source rock for the oils of the Williston Basin. Numerous studies have examined the geochemistry and hydrocarbon potential of the Bakken shales (Dow, 1974; Williams, 1974; Webster, 1982; 1984; Schmoker and Hester, 1983; Price *et al.*, 1984). However, recent information indicates that a large amount of oil that has been produced and previously thought to have been sourced by the Bakken Formation was, in fact, generated within the Lodgepole Formation (Osadetz and Snowdon, in review; Osadetz *et al.*, in review).

The hydrocarbon potential of the Bakken Formation has been estimated to be at least 98 billion barrels (Webster, 1982; 1984). Based on the recent findings of the Geological Survey of Canada, a larger amount of undiscovered Bakken oil exists somewhere within the Williston Basin. Research indicates that this oil is probably stratigraphically trapped within a narrow zone that surrounds and includes the Bakken Formation.

Conventional Bakken production in North Dakota is restricted to areas where the Bakken is a thermally mature source rock. However, well beyond there are fields in Saskatchewan and Manitoba which produce from a sandstone within the middle member of the Bakken. This suggests that there has been an outward lateral migration of Bakken oil from the central portion of the Williston Basin. Recent focus on the Bakken Formation has been directed toward the development of and exploration for oil in the Bakken shales of North Dakota and Montana. This study will examine the overall stratigraphy and petroleum potential of the Bakken Formation in North Dakota, Manitoba, and Saskatchewan (Fig. 1).

GEOLOGICAL SETTING

The Williston Basin is an intracratonic, structural, and sedimentary basin located at the western edge of the Canadian Shield. It occupies portions of North Dakota, South Dakota, Montana, Saskatchewan, and Manitoba (Fig. 1). Gerhard *et al.* (1982) related the formation of the basin and other major structural features, including the Nesson, Cedar Creek, and Billings anticlines, to a major directional change in the structure of the Rocky Mountain belt.

Nearly continuous sedimentation has occurred in the North Dakota portion of the Williston Basin. This is represented by over 4,878 m (16,000 ft) of sediments representing Cambrian through Tertiary time. Basin sedimentation is characterized by cyclical transgressions and regressions with the repeated deposition of carbonates and clastics. Paleozoic strata are dominated by carbonates, whereas Mesozoic and Cenozoic strata consist mainly of clastic rocks.

Initial sedimentation occurred over an irregular Precambrian surface. Repeated clastic/carbonate sequences interrupted by major erosional events are evidenced by strata from Cambrian to Devonian time. The Devonian period marked a reorientation of the seaway to the north as a result of activity along the Transcontinental arch. Devonian sediments show the repeated transgressional-regressional cycles related to Elk Point basin deposition. A reorientation of the seaways occurred again during the Mississippian when the basin opened to the west through the central Montana trough. Terrestrial, marginal marine, and evaporite sediments are represented by Pennsylvanian through Triassic strata. The last sequence of marine strata was deposited during the Jurassic and Cretaceous periods.

A major angular unconformity separates the Paleozoic from the Mesozoic strata and probably represents one or more periods of erosion that occurred from Late Mississippian to Early Jurassic time. During this interval, Paleozoic strata in the northeastern part of the basin were uplifted and differentially eroded, while strata in the southern portion of the basin were relatively unaffected (McCabe, 1959). Successively older Paleozoic strata were progressively truncated toward the basin margin. Deposition resumed during Mesozoic time, when a thick sequence of Jurassic and Cretaceous strata was deposited on the eroded Paleozoic surface.

Within the Paleozoic itself, an unconformity separates Devonian and Mississippian strata and represents uplift and erosion which occurred from Late Devonian to Early Mississippian time. During that interval, Devonian strata were uplifted and exposed along the basin margins, while deposition continued in the deeper portions of the basin. Mississippian sediments were later deposited on the eroded Devonian surface (Sandberg, 1964).

In North Dakota, the Bakken Formation is a thin, clastic unit which straddles the Devonian-Mississippian boundary. It occurs solely in the subsurface and consists of three informal units, referred to as the lower, middle, and upper members in this report. The formation reaches a maximum thickness of 46 m (150 ft) in the central portion of the basin (Fig. 2).

The Bakken Formation subcrops in Manitoba along the northeastern portions of the study area, where it is unconformably overlain by the "Red Beds" of the Amaranth Formation (Jurassic).

Southeastern Saskatchewan is situated on the northern shelf adjacent to the Williston Basin (Christopher, 1961). Sediments associated with the margin form a southerly-dipping wedge that thins towards the northern erosional edge. The strong southerly tilt superimposed on these sediments occurred between Mississippian and Jurassic time and is probably related to basin subsidence.

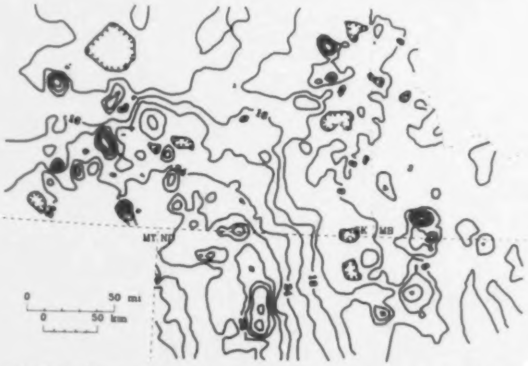


Figure 2: Isopach map of the Bakken Formation for the study area. Contour interval is 4 m (13 ft). Dashed line in northeastern Manitoba is the erosional limit of the Bakken Formation.

GENERAL STRATIGRAPHY

A persistent sequence of black shales and grey siltstones was recognized at the base of the Madison Group (Mississippian) limestones early in the exploration of the Williston Basin. These beds were referred to by a variety of names, including Kinderhook, Englewood, and Exshaw (Fuller, 1956). In an attempt to eliminate the confusion, a formal name, the Bakken, was introduced into the literature in 1953 by the Williston Basin Nomenclature Committee of the Saskatchewan Society of Petroleum Geologists and the Rocky Mountain Section of the American Association of Petroleum Geologists. No formal definition for the formation was presented at that time.

Nordquist (1953) formally defined and described the Bakken Formation for strata occurring at depths from 2,931 to 2,963 m (9,615 to 9,720 ft) in the Amerada Petroleum Corporation - H.O. Bakken #1 deep test (Fig. 3). The well is located in CSWNW Sec. 12, T.157N., R.95W. in Williams County, North Dakota. As based on samples, the formation is 32 m (105 ft) thick at its type section and consists of two black, organic-rich shales which are separated by a light grey to grey-brown, very fine-grained, calcareous sandstone.

Kume (1963) modified Nordquist's original description of the Bakken type section. He re-evaluated the thicknesses of the members, modified the sample descriptions for the lower and middle member, and designated a standard subsurface reference section. The standard reference section is located over the interval from 3,059 to 3,078 m (10,035 to 10,095 ft) in the Socony Vacuum Oil Company - C. Dvorak #1 well (SENE Sec. 6, T.141N., R.94.W. Dunn County, North Dakota). This well was selected because it was continuously cored from the lower portion of the Madison Group (Mississippian) into the Birdbear Formation (Devonian).

The Bakken Formation is formally recognized in southwestern Manitoba, southern Saskatchewan, and eastern Alberta (Fuller, 1956; MacDonald, 1956; McCabe, 1959; Christopher, 1961; 1962; Carlson and LeFever, 1987). In southwestern Alberta, the equivalent strata are divided into two formations: the basal black shale and middle siltstone member are included in the Exshaw Formation; and, the upper black shale is included with the basal shales of the Banff Formation (MacDonald, 1956). Usage of Exshaw and Banff formations is carried into northern Alberta and northern British Columbia. Correlations in northern Alberta and British Columbia become more difficult as the shale become lithologically similar to the underlying and overlying beds (Macauley *et al.*, 1961).

The Bakken Formation is an easily recognizable sequence on wireline logs in the subsurface of the Williston Basin. As previously stated, it comprises three members, a lower and an upper shale and a lithologically variable middle member. Readily apparent on wireline logs, the shales have abnormally high gamma-ray readings (>200 API), low resistivity (ohm-m) readings in the shallower portion of the basin, high resistivity readings in the deeper portion of the basin, and high transit times (80 to 120 microsec/ft) (Meissner, 1978; Webster, 1982; 1984). The middle member has normal wireline log characteristics for clastics and carbonates and becomes difficult to distinguish from the underlying Three Forks Formation when the lower shale is absent.

The three members of the Bakken Formation exhibit an onlapping relationship; each successively higher member has a greater areal distribution. The formation overlies the Three Forks Formation (Upper Devonian) in North Dakota and Manitoba, and the Upper Devonian Torquay and Big Valley formations in Saskatchewan. In turn, the formation is overlain by the Lodgepole Formation (Lower Mississippian) throughout North Dakota and Manitoba and by the Souris Valley Beds (Lower Mississippian) in Saskatchewan.

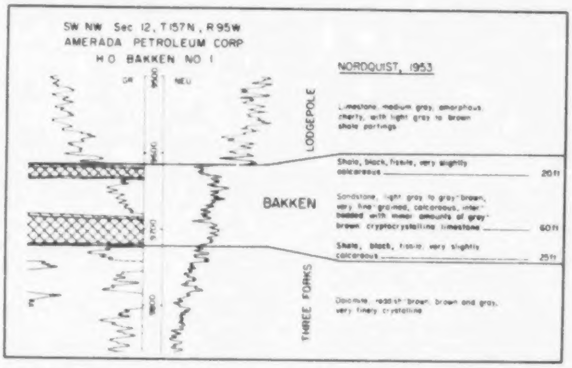

SW NW Sec 12, T157N, R95W
AMERADA PETROLEUM CORP
H O BAKKEN NO 1
GR
NEU
9500
9600
9700
9800
LODGEPOLE
BAKKEN
THREE FORKS
NORDQUIST, 1953
Limestone, medium gray, amorphous, cherty, with light gray to brown shale partings
Shale, black, fissile, very slightly calcareous
20ft
Sandstone, light gray to gray-brown, very fine-grained, calcareous, interbedded with minor amounts of gray-brown cryptocrystalline limestone
60ft
Shale, black, fissile, very slightly calcareous
25ft
Dolomite, reddish-brown, brown and gray, very finely crystalline

Figure 3: Type log for the Bakken Formation from Amerada Petroleum Corporation - H.O. Bakken #1 deep test. Lithologic description is from Nordquist (1953).

BAKKEN FORMATION

Lower Member

The lithology of the lower member of the Bakken Formation is uniform over the entire study area. It consists of a dark grey to brownish-black to black, non-calcareous, fissile, slightly to highly organic-rich shale. The colour of the shale varies depending on the amount of silt versus clay versus carbon present in the rock. Lesser amounts of siltstone, limestone, and sandstone occur within the lower member in North Dakota. The shale is finely laminated to massive and can be hard or soft ("wax-like").

In Saskatchewan, Fuller (1956) described a 2.5 cm (1 in) thick pebble bed at the base of the Bakken Formation. Constituents include dolostone fragments, occasional broken conodonts, and chert in a grey siltstone matrix. Phosphatic fragments are also occasionally present. Fuller suggested that the pebble bed represented the basal conglomerate of the Bakken Formation. Hayes (1984) described a thin, 2.5 cm (1 in), coarse, lag sandstone bed at the base of the lower member in North Dakota. The sandstone bed contains fragments of conodonts, pyritized clasts, phosphatic particles, fish bones, subangular quartz sand and silt, chert, carbonate debris, and rounded pebbles all floating in a shale matrix.

Fossils identified by Hayes (1984) in the lower member in North Dakota include algal plant spores (*Tasmanites*, sp.), conodonts, inarticulate brachiopods (*Lingula* sp., *Orbiculoidea* sp.), fish teeth, bones, and scales, conchostracans, rare ostracods, woody plant fragments, and sponge spicules. Fossils identified in Saskatchewan include *Lingula*, conodonts, spores or spore-like remains including *Tasmanites* (Fuller, 1956; Christopher, 1961).

The lower shale is organic-rich in the deeper portion of the basin and is considered to be a mature source rock. The organic material is distributed evenly throughout the member. Pyrite is also abundant, and occurs in thin, wispy laminae, in lenses or nodules, or disseminated throughout the interval. Fractures can also be present, can be irregular and blocky or smooth and conchoidal in the more siliceous sections. Occasionally, the fractures are healed with calcite or pyrite. Resinous or vitreous pods of carbonaceous material ("dead oil") are present along some of the fractures which occur parallel or subparallel to bedding.

The lower member reaches a maximum thickness of 17 m (55 ft) in North Dakota with a well-defined depocenter immediately east of the Nesson anticline. Thicknesses in Saskatchewan reported by Christopher (1961; 1962) for the lower Bakken shale member generally range from 6 to 8 m (20 to 25 ft). A local exception is the Sohio-Leakville #1 (4-11-14-26 W2M) where the lower Bakken shale member reaches a thickness of 13 m (43 ft). In Manitoba, the Lower Member is restricted to the Waskada area (Townships 1 and 2; Ranges 25 and 26 WPM) and is a maximum 13 m (43 ft) thick (McCabe, 1959; Martinuik, 1988).

Middle Member

The lithology of the middle member is highly variable in North Dakota. It consists of a light grey to medium dark grey interbedded sequence of siltstones and sandstones. Lesser amounts of shale, dolostones, and limestones rich in silt, sand, and oolites may also be present (Webster, 1982; Hayes, 1984; Thrasher, 1985). The siltstones and sandstones are massive or coarsely bedded with an occasional trough or planar cross-bed set. Much of the sequence is well sorted. Bioturbation commonly disturbs the bedding, especially in the more argillaceous portions. Features indicative of soft sediment deformation, including microfaults and flow structures, are also present. Cements include calcite, dolomite, silica, and pyrite. Insoluble residues comprise quartz silt and sand, feldspar, chert, glauconite, and silicified and pyritized fossils.

Fossils within the middle member include articulate and less common inarticulate brachiopods, pelmatozoan fragments, gastropods, and various trace fossils. Conodonts, plant spores, and ostracods were also observed but are considered to be rare (Hayes, 1984).

In Manitoba, the Middle Member consists primarily of a light greenish-grey to reddish- and purplish-grey siltstone to very fine- to medium-grained sandstone (Martiniuk, 1988). Much of the sequence is dolomitic and argillaceous. Mineralogically, the sandstones consist predominantly of quartz with minor amounts of feldspar. Coarser grain sizes are generally subrounded. The matrix is a dolomite or dolomitic shale, depending on the degree of sorting. Grain size increases in the northern and eastern portion of the study area.

McCabe (1959) reported that often the Middle Member is massive or finely laminated. Cross-bedding and contorted bedding were also noted. Several intraformational breccias and conglomeratic zones observed near the base of the unit probably represent diastemic and/or erosional breaks within the member.

In Saskatchewan, Christopher (1961; 1962) divided the middle sandstone member of the Bakken Formation into two units, A and B. Unit A, the lowermost unit, consists of a massive, calcareous, greenish-grey, pyritiferous, fossiliferous (primarily brachiopods), very fine-grained to silty sandstone that is present in the northwestern, western and southern portion of Saskatchewan. Unit B consists of an interfingering shale/sandstone sequence, designated B1 through B4. The shale units, B1 and B3, occur west of the second meridian. East of the second meridian, the two sandstone units, B2 and B4, come in contact to constitute the middle Bakken sandstone member. The uppermost unit, B4, represents a massive, fossiliferous, silty sandstone or coarse siltstone. This unit, in turn, is underlain by B2, a fine- to medium-grained, friable sandstone which changes westward into a well cemented, calcareous, medium-grained sandstone and oolitic calcarenite. Unit B2 exhibits well developed cross-bedding, ripple marks, current scour marks, and numerous diastems.

The middle member attains a maximum thickness of 27 m (87 ft) in North Dakota. Its depocenter, like the lower member, is well developed and is located just to the east of the Nesson anticline. In Manitoba, the Middle Member averages 4 m (13 ft) in thickness and reaches a maximum thickness of 16 m (52 ft) in the Waskada area. In the study area of Saskatchewan, the middle Bakken sandstone member reaches a maximum thickness of 18 m (60 ft) in the Herald and Torquay embayments (Christopher, 1961).

Upper Member

The upper member of the Bakken Formation is lithologically similar to the lower member and is also uniform throughout the study area (Fuller, 1956; McCabe, 1959; Christopher, 1961; 1962; Webster, 1982; Hayes, 1984; Martiniuk, 1988). It consists of dark grey to brownish-black to black, fissile, noncalcareous, carbonaceous and bituminous shale. It is more fossiliferous than the lower shale. Christopher (1961; 1962) reported abundant conodonts and chonetid brachiopods. Hayes (1984) noted the presence of other fossils including fish teeth, bones, and scales, rare woody plant fragments, inarticulate brachiopods, and spores in addition to conodonts. He also noted that one well had a well developed layer of reworked, pyritized, articulate brachiopods.

The upper member differs from the lower shale in North Dakota with the absence of crystalline limestones and greenish-grey shale beds and an increase in the organic content. Fractures are also present in the upper member. Thin, poorly defined laminae of silt-sized quartz and carbonate grains occur throughout. Small sandstone lag deposits were also described in the upper member by Hayes (1984). These deposits are rich in conodonts, fish bones and teeth, and phosphatic particles and occur at several zones within the shale.

The upper member reaches a maximum thickness of 9 m (28 ft) in North Dakota. The depocenter is poorly defined. In Saskatchewan, the shale is a maximum 4 m (12 ft) thick, averaging 1.2 m (4 ft). A uniform thickness of 2 m (7 ft) for Manitoba is reported by Martiniuk (1988). In the Waskada area of Manitoba, the Upper Member attains a maximum thickness of 18 m (59 ft).

Depositional Environments

A variety of environments have been suggested for the origin of the two black shales of the Bakken Formation. Workers agree that the environments of deposition were similar for the lower and upper shales and that conditions were uniform over the area. However, a wide range in ideas is possible when conditions conducive for shale deposition vary from open marine to terrestrial swamp.

Fuller (1956) suggested that the black shales were deposited in a vast swamp resulting from the retreat of Devonian seas. Iron oxides and sulfides reworked from Qu'Appelle Group sediments were reduced to form the iron sulfides of the Bakken muds. A transgression with the resulting incursion of seawater eroded the stable margins and formed the arenaceous beds of the middle member. A final seaward retreat of waters was followed with the return of conditions responsible for the formation of the black shales.

McCabe (1959) also suggested a marine swamp for the formation of the black shales. Restriction of circulation was due to prolific organic development. During the deposition of the Middle Member, the swamp was drowned or flooded by an influx of shallow marine clastics derived from older Paleozoic or Precambrian rocks. Stagnant, swamp conditions were reestablished for the deposition of the Upper Member. McCabe stated that the underlying Three Forks, the overlying Lodgepole, and the Middle Member were deposited in shallow marine conditions; therefore the shales were also probably deposited in shallow marine to terrestrial environment.

Christopher (1961; 1962) attributed the environment of deposition of the Bakken shales to a restriction of free flow water in a shallow sea due to sags and swells. Periodically fetid, the study area covering Saskatchewan converted to tidal flats and broad swamps. Thick black muds accumulated locally in sinkholes. Deposition of the middle Bakken sandstone member occurred in response to the southwesterly tilt of the Saskatchewan portion of the study area. Clastic sources of silt and clay were introduced into the area from the northeast and east and filled in basinal areas to the southeast, west, northwest and north. The initial transgression of the middle Bakken seas resulted in the rapid deposition of clays and silts and the burial and preservation of abundant brachiopods. After the initial deposition of Christopher's "A" unit, the eastern area began to subside and the sea spread over the platform area of Saskatchewan to the flat shelf area to the east and into Manitoba. This strandline and erosion of the Devonian extended into Manitoba. The deposition of the "B" unit followed with fine- to medium-grained sands intermittently laid down on a broad shelf in shallow water of variable current activity. Grey-black muds interfingered with the sands reflecting a "sea floor of flats, shoals, and broad hollows traversed by weak, shifting currents" (Christopher, 1962, p. 76). Sedimentation slowed and returned to swamp/lagoonal conditions, under which the upper Bakken shales were deposited. Bakken sedimentation was terminated by a major transgression under which the argillaceous limestones of the Madison Group were deposited.

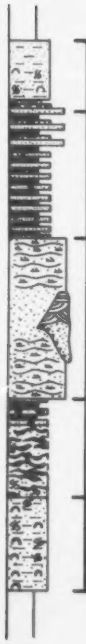
Webster (1982) and Lineback and Davidson (1982) suggested a stratified water column for the deposition of the black shales. Evidence presented by Webster in support of this idea includes planar and thin laminations within the shale resulting from quiet water deposition below wave base; presence of planktonic algal spores (*Tasmanites* sp.), fish remains, cephalopods, ostracods, conodonts, and inarticulate brachiopods; high amounts of organic and pyritic material suggesting anaerobic conditions; uniformity over a large area; and the predominance of amorphous-sapropelic organic matter over terrestrial material. A temperate climate could account for the lack of mixing between the bottom and surface waters. The lithology and fossil content of the middle member suggests a return to aerobic marine conditions with moderate current activity.

Age

Early workers assigned a Mississippian age to the Bakken Formation based on lithology and limited paleontological data (Nordquist, 1953; Thomas, 1954; Fuller, 1956). McCabe (1959) included the Bakken as the basal formation of the Mississippian system. This assignment was based on the absence of the Lower Member of the Bakken over most of Manitoba and the evidence of erosion at the top of the Three Forks Formation, both indicating a possible time break. Martiniuk (1988) also placed the Bakken in Manitoba as the basal unit of the Mississippian sequence with the systemic boundary inferred at the unconformity between the Bakken and underlying Three Forks Formation.

Table 1: Composite lithofacies descriptions and correlations for the middle member of the Bakken Formation for the study area. Christopher (1961) divided the middle member into two units, A and B. Unit B was then further subdivided into B1, B2, B3, and B4. Thrasher (1985) and Karma and Parslow (1989) used macrofossils and lithology to subdivide the middle member into three units. This study incorporates both divisions of the middle member.

LITHOFACIES



	This Paper			Thrasher, 1985 Karma & Parslow, 1989	Christopher, 1961
UPPER MEMBER	SK	ND	MB	SK,ND	SK
SILTSTONE, massive, dense, mottled, dolomitic, argillaceous, grey-green, fossiliferous, disseminated pyrite, rhythmites up to 15 cm thick in lower half of section (occassionally fossil-rich), slightly bioturbated, contact with upper member sharp.	C	7	4 3	3	B4
PARALLEL INTERBEDS OF DARK GREY SHALE AND BUFF SILTY SANDSTONE, moderately bioturbated, vertical burrows, calcareous, disseminated pyrite, overall coarsening-upward, flame and load structures at base of coarse laminae, rhythmites up to 10 cm thick, upper half may display trough cross-bedded sandstone beds, gradational lower contact with underlying unit.	B3	6 5		2	B3
SANDSTONE, tripartite division with upper and lower third wavy and flaser bedded silty sandstone gradational to and from the middle coarse-grained sandstone which may be massive and/or bedded (trough and tabular cross-bedding, inclined and horizontal laminae, and swash cross-stratification) with pebble and fossil-rich lags (shale clasts up to 4 cm diameter where B2 overlies the lower member, and feldspar clasts). Mainly quartzose with minor feldspar and heavy minerals, disatems, rhythmites, few brachiopods, disseminated pyrite, buff to green, calcareous, slight to no bioturbation.	B2	4 3	2 1		B2
PARALLEL INTERBEDS OF DARK GREY SHALE AND BUFF SILTY SANDSTONE, moderate to very strong bioturbation disrupting laminae, dolomitic becoming calcareous with depth, disseminated pyrite, fossiliferous, lower contact gradational, upper contact gradational or erosive where channelling, grey-green.	B1	2			B1
SILTSTONE, massive, dense, mottled, very calcareous, argillaceous, grey-green, highly fossiliferous, random orientation of fossils, disseminated pyrite, lower contact may be either gradational over several centimeters or erosive.	A	1		1	A

LOWER MEMBER

Christopher (1961; 1962) placed the systemic boundary at the base of the middle Bakken sandstone member in Saskatchewan. This decision was based on the lack of Kinderhookian age fossils in the lower Bakken shale member. Earliest recognizable Kinderhookian age fossils were described by Brindle (1960) in the middle sandstone member of the Bakken in Saskatchewan.

Detailed paleontological studies have been conducted on the Bakken Formation in North Dakota by Hayes (1984), Thrasher (1985), and Holland *et al.*, (1987). These studies suggest that the Bakken Formation is Devonian and Mississippian in age. Using conodont fauna from the shales, Hayes (1984) suggested that the systemic boundary occurred at or near the contact between the middle and upper members. The upper member has characteristically lower Mississippian conodonts that are not the earliest Kinderhookian. This suggests that there is a hiatus between the two that may coincide with a regionally extensive unconformity. Thrasher (1985) examined macrofossils of the middle member and concluded that the systemic boundary appears to be within the middle member.

DETAILED STRATIGRAPHY

The following is a detailed description of the lithofacies of the Bakken Formation based on data derived from the examination of core in North Dakota, Saskatchewan, and Manitoba. Composite lithofacies descriptions of the middle member, showing the correlations of the lithofacies identified, are shown in Table 1. The regional fence diagram (Fig. 4) shows the distribution of the various lithofacies over the study area.

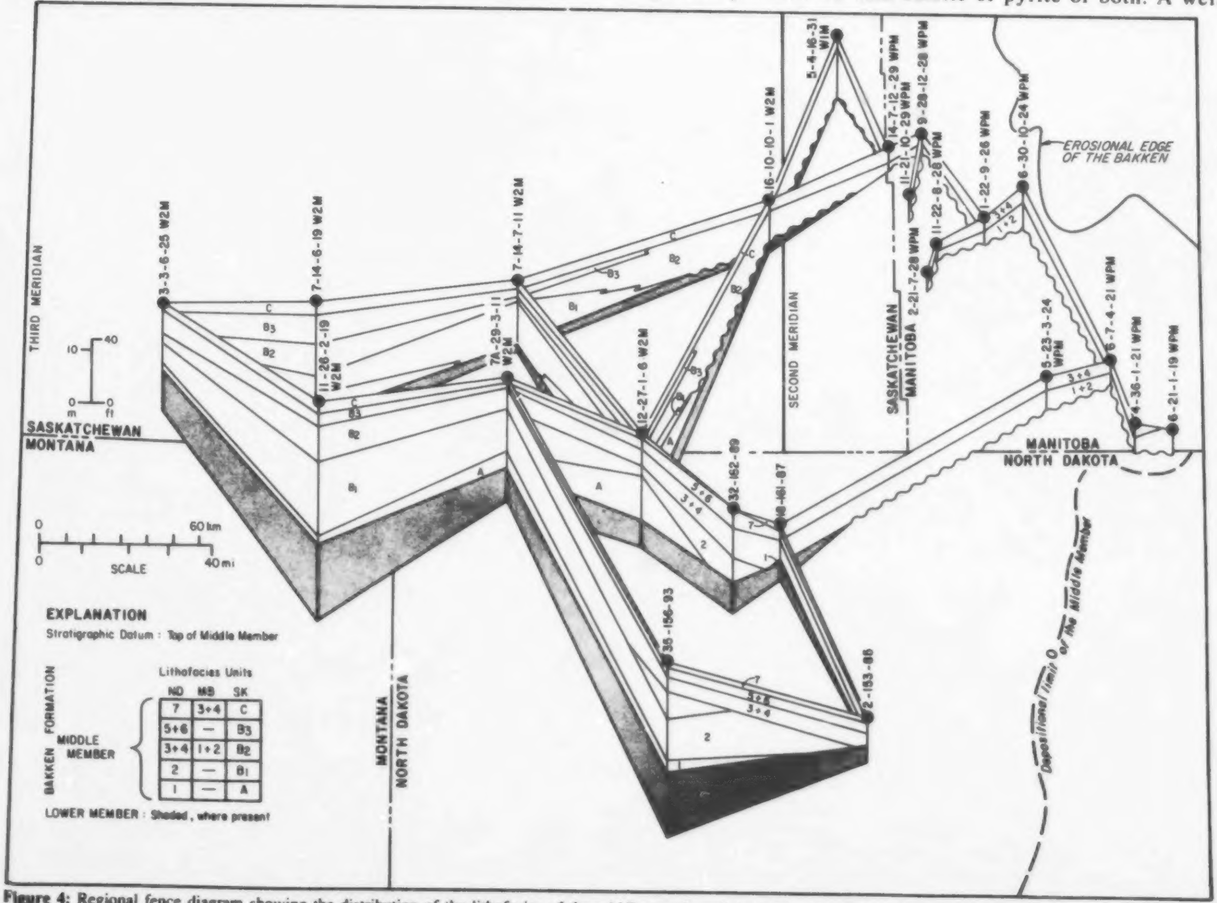
NORTH DAKOTA

Bakken Formation: Lower Member

The lower member of the Bakken Formation overlies the greenish-grey, argillaceous burrowed siltstone of the Three Forks Formation. The lowermost contact was observed only in one core and was irregular and sharp. The contact was overlain by a bed containing small clasts of bitumen and fossil fragments.

The lower member consists of a dark brown-black to black, organic-rich shale. It can be massive and competent or fissile. Pyrite is generally disseminated throughout the shale; it also occurs in lenses, as nodules, or in laminations. In addition to pyrite grains, laminations formed by the alignment of silt-sized calcite grains are also present. Conodonts, ostracods, cephalopods, conchostracans, and brachiopods constitute the fossil assemblage of the shale.

A variety of fractures are associated with the lower member. Thin, hair-like, vertical fractures occur in the more massive zones. Conchoidal fractures are present when the lithology show an increase in silt content. Open bedding plane fractures with residual oil staining are also present. Closed vertical fractures are generally cemented with calcite or pyrite or both. A well



ND	MB	SK
7	3+4	C
5+6	—	B3
3+4	1+2	B2
2	—	B1
1	—	A

Figure 4: Regional fence diagram showing the distribution of the lithofacies of the middle member of the Bakken Formation. The lower member is shaded where present. Lithofacies units with their appropriate number or letter designation are shown.

developed breccia zone with calcite cement occurs in one well (SWNE Sec. 26, T.156N., R.91W.). Limestone interbeds are also present within this well. Many of the wells show small-scale deformed fractures, all of which are cemented with calcite or pyrite or both and are associated with laminations of silt-sized pyrite and calcite grains. Sizes range from 0.6 cm (.25 in) to several centimetres. Christopher (1961) suggested that these "vertical fissures" may have originated as mudcracks.

The contact between the lower member and the middle member is irregular and sharp or gradational. Beds containing small clasts and fossil fragments commonly overlie the sharp, irregular contacts. The gradational contacts show either a gradual decrease in argillaceous content to the siltstone/very fine- grained sandstone lithology of the middle member or a sequence of alternating shale and siltstone/sandstone beds that decrease in argillaceous content vertically. The gradational contacts are restricted to the northernmost wells in North Dakota.

Bakken Formation: Middle Member

Seven lithofacies were observed in the cores of the middle member in northern North Dakota. These facies are correlative with those recognized in cores in Manitoba and Saskatchewan (Table 1 and Fig. 4). Data are derived from Bakken core examined in 12 North Dakota wells. The seven lithofacies recognized in North Dakota, in ascending order, are as follows:

Lithofacies Unit 1 consists of a light grey, greenish-grey, or brownish-grey argillaceous siltstone. It is massive, cemented with calcite, and has scattered pyrite nodules, and fossils. Fossils generally include crinoids and brachiopods. One well shows abundant small brachiopods above the contact with the lower shale. Burrowing is locally present. The lithofacies range in thickness from 0.5 to 1.8 m (1.5 to 6 ft). Intergranular porosity and permeability for unit 1 in one well was determined to be 3.7 percent and 0.01 md, respectively.

Lithofacies Unit 2 is greenish-grey to brownish-grey, argillaceous siltstone or sandy siltstone to a brownish-grey, very fine-grained sandstone. The unit ranges up to 10 m (33 ft) in thickness. Small-scale clay drapes are present throughout, as well as burrows filled with pyrite. In one well (NENW Sec. 35, T.156N., R.93W.), where the unit is 8 m (27 ft) thick, burrowing is restricted to the upper 2 m (7 ft) and basal 1.5 m (5 ft). Scattered fossils, crinoids and brachiopods, are present throughout. Identifiable brachiopods include *Chonetes ornatus* and *Rhytiophora arcuatus*. Fossils are generally partially or completely replaced with pyrite.

Glauconite was noted in three wells over a restricted interval ranging from 0.6 to 1.2 m (2 to 4 ft). The glauconite is often associated with burrows. Hematite staining was also noted in the basal portion of one well. Cement, where present, consists of calcite and appears where the interval is coarser grained. Porosities range from 1.2 to 8.4 percent, averaging 5.5 percent. Permeabilities range from <0.01 to 0.41 md, averaging 0.22 md. Unit 2 was oil productive in one well with treatment.

One well, toward the central portion of the basin (NENW Sec. 35, T.156N., R.93W.), has an additional 5 m (16 ft) section between units 2 and 3 which is not present in the other wells (Fig. 5). The basal bed consists of a laminated, slightly argillaceous, oil stained, sandy siltstone, which grades upward into a medium brownish-grey, massive siltstone; an interbedded sequence of very fine-grained sandstone and claystone; and, a medium greyish-brown, uniformly laminated, oil-stained sequence of siltstone and very fine-grained sandstone. Individual sandstone beds are moderately well sorted and some are lensoidal.

Lithofacies unit 3 is the sandstone bed which is productive in Canada (Table 1). In North Dakota, the bed is prominent both in core and on wireline logs (Fig. 6) and easily mapped (Fig. 7). It is a medium grey, dark grey or greyish-tan, very fine- to fine-grained sandstone. Occasionally, the grain size increases into the upper medium range. The sandstone is moderately well sorted to well sorted and can be poorly sorted locally. Grains are generally well-rounded to rounded. Some of the finer grained material is subrounded to subangular. The sandstone consists predominantly of quartz with occasional feldspar and heavy minerals. Glauconite was observed in one well. Coarser grained material consists locally of rounded to subrounded ooid and crinoid fragments. In the wells where these sediments occur quartz is rare. Pyrite grains or nodules are scattered throughout. The sandstone is cemented generally by calcite and occasionally by pyrite.

Unit 3 consists commonly of an alternating sequence of massive, cross-bedded, and thinly laminated beds. The cross-bedded and massive beds are generally coarser grained. The upper portion may have multiple fining-upward sequences, whereas the lower portion may display reverse grading. Other features observed include load or channel forms grading upward into fine-grained material, and calcite-filled fractures. Soft-sediment deformation was observed in the middle of the unit in a few cores.

Unit 3 ranges in thickness from 0.2 to 4.6 m (0.5 to 15 ft), averaging 2 m (8 ft). The contacts between Unit 3 and the overlying and underlying units are sharp.

Measured porosities range from 0.09 to 12.1 percent, and average 7 percent. Permeabilities range from 0.02 to 5.8 md, and average 0.6 md. Oil staining appears to be restricted to the very fine-grained, thinly laminated quartz sandstone beds.

Lithofacies Unit 4 comprises interbedded light to medium grey, argillaceous to sandy siltstones and brownish-grey, very fine-grained sandstones and local claystone. The sequence is commonly laminated. Laminations are generally irregular or wavy bedded, or highly disturbed due to soft sediment deformation. Cross-ripple laminations and reverse grading are also present. There are occasional lenses of coarser-grained material in which the intergranular pores are filled with pyrite. Some sections also show extensive pyrite-infilled burrowing. A 0.3 m (1 ft) thick disturbed zone, due to soft-sediment deformation, is commonly present at the base of the unit.

Unit 4 reaches a maximum of 2 m (6.5 ft) in thickness. The unit is generally not cemented. Cement, where present, consists of dolomite.

Measured permeabilities are low (<0.01 to 0.15 md). Overall porosities are higher than the lower lithofacies, and range from 3.1 to 15.8 percent, averaging 8 percent.

Lithofacies Unit 5 consists of an alternating sequence of medium grey, argillaceous siltstones, light to medium grey, very fine-grained sandstones, and dark grey shale. It is thinly laminated. Laminations are either parallel or slightly undulatory and range in thickness from 1 to 5 mm (.04 to 0.2 in), with grain size increasing slightly in the thicker laminations. The unit is locally cemented with dolomite. Several of the wells exhibit soft-sediment deformation as contorted bedding either at the top or bottom of the interval or as small folds and faults (post-lithification) within.

Thicknesses of unit 5 ranges from 0.6 to 1.1 m (2 to 3.5 ft). Permeabilities range from <0.01 to 0.56 md, averaging 0.2 md. Porosities range from 3.8 to 12.4 percent, and average 7 percent.

Lithofacies Unit 6 consists of an alternating sequence of medium to light grey siltstone, dark grey claystone to brown/black shale; and, tan to light brownish-grey, very fine-grained sandstone, siltstone, and sandstone beds that commonly display planar or cross-ripple laminations. The sequence occurs as beds or lenses and is usually disrupted due to burrowing. The basal portion of the unit (0.3 to 0.9 m; 1 to 3 ft) consists predominantly of a thinly laminated, very fine-grained sandstone and siltstone with abundant burrows. Argillaceous content varies locally and generally increases upward within the section. The unit is locally cemented with dolomite.

Unit 6 varies in thickness from 0.9 to 3.2 m (3 to 10.5 ft), and averages 2 m (8 ft). Porosities range from 1.1 to 13.8 percent, averaging 5.5 percent. Permeabilities range from <0.01 to 97 md.

Lithofacies Unit 7 is a medium to light grey, massive to wispy laminated siltstone. Locally, the unit becomes greenish-grey where the argillaceous content increases and is generally cemented with calcite. Dolomite cement is present over a restricted interval in two cores. Brachiopods occur throughout and include *Chonetes gregarious*, *Chonetes ornatus*, and *Composita*. Cores from wells in the central portion of the basin also include bryozoan and crinoid fragments. Many of the fossils are partially or completely replaced with pyrite.

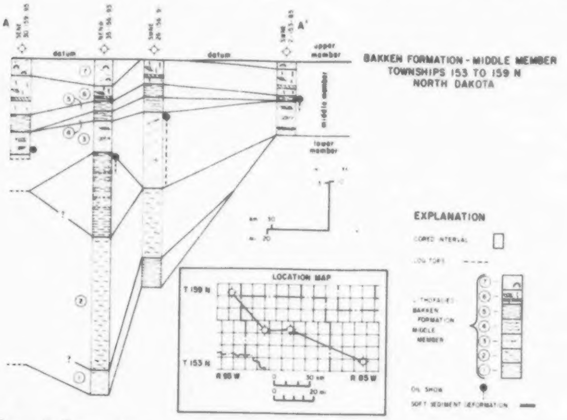


Figure 5: Fence diagram showing the correlation of lithofacies units of the middle member of the Bakken Formation from Townships 153 to 159 N, North Dakota. The lithofacies represented were described from cores. Oil shows are indicated over the interval where present.

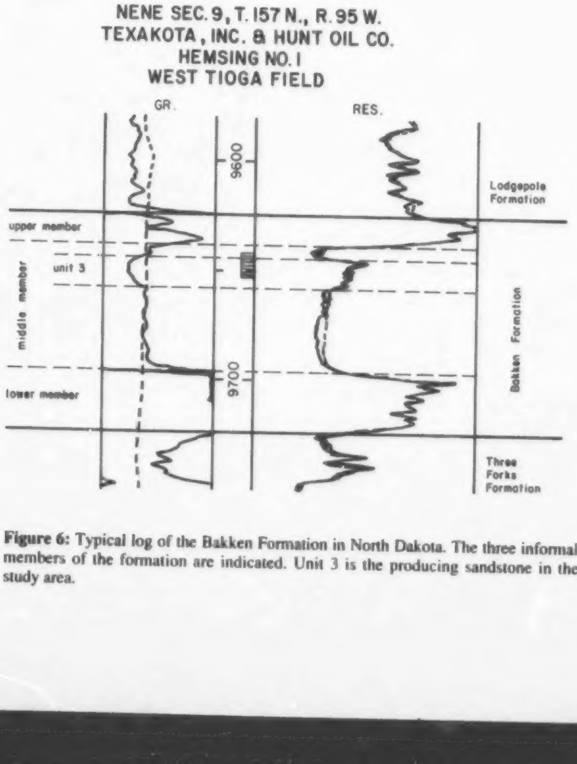


Figure 6: Typical log of the Bakken Formation in North Dakota. The three informal members of the formation are indicated. Unit 3 is the producing sandstone in the study area.

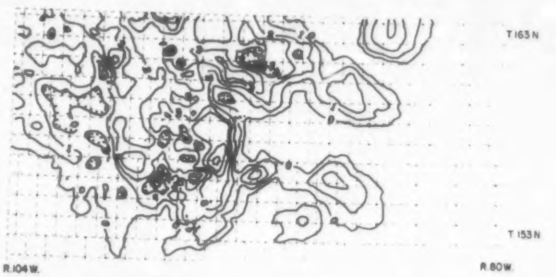


Figure 7: Isopach map showing the distribution of lithofacies unit 3, the producing sandstone, throughout northwestern North Dakota. Contour interval is 1 m (3 ft). Dashed lines are townships and are 10 km (6 mi) on a side.

Thin beds of tan to light grey, very fine-grained sandstone occur toward the bottom of unit 7 and are commonly ripple cross-laminated. Beneath one of these cross-bed sets is an argillaceous layer, rich in brachiopods. This layer was observed in several wells.

Unit 7 ranges from 0.6 to 1.8 m (2 to 6 ft) in thickness. Permeabilities range from <0.01 to 109 md. Porosities range from 0.02 to 12.8 percent. Accumulations of pyrite occur near the upper contact with the upper member.

Bakken Formation: Upper Member

The upper member of the Bakken Formation consists of an organic-rich, slightly calcareous, massive to fissile, dark brown-black to black shale. The alignment of calcite and/or pyrite grains forms thin wispy laminations. Pyrite is disseminated throughout but also concentrated into lenses, nodules, and laminations. Small-scale fractures, observed in the lower member, are also present in the upper member. These fractures are filled with calcite and pyrite. Fossils observed within the shale include abundant conodonts and rare brachiopods.

The contact between the middle member and upper member appears to be sharp and irregular or gradational. In wells where the contact is sharp, a thin lag zone with pyrite and bitumen occurs along the surface. The gradational contacts exhibit a gradual upward increase in argillaceous content.

SOUTHWESTERN MANITOBA

Three Forks Formation

The Three Forks Formation is unconformably overlain by the Middle Member of the Bakken Formation throughout most of southwestern Manitoba. The Three Forks generally consists of light grey-green, blocky or massive, waxy, silty shale with disseminated pyrite. It may also occur at the Bakken contact as a buff, light grey-green or purple, silty, argillaceous dolostone; a purple, green and brown varicoloured or mottled, massive, dolomitic shale; or as a medium-grey, argillaceous, silty shale or interbedded siltstone and shale. It is difficult to distinguish Bakken from Three Forks lithologies in some areas.

Bakken Formation: Middle Member

Four lithofacies were identified within the Middle Member in southwestern Manitoba. These facies are correlative with those recognized in Saskatchewan and North Dakota (Table 1; Fig. 4). Data were derived from Bakken core examined in 32 southwestern Manitoba wells. The four lithofacies, in ascending order, are identified as follows:

Lithofacies Unit 1 is a buff to light grey or tan, very fine- to fine-grained sandstone. The unit may be massive to friable, slightly wavy bedded, or thinly laminated. Small-scale ripple cross-laminations are occasionally present near the top. Low-angle cross-bedding is also present in some cores. Wavy bedding occurs as 0.5 to 1 cm (0.2 to 0.4 in) thick lenses. Laminae generally consist of dark grey shale or green shale. Green shale laminae are often concentrated near the lower contact with the Three Forks Formation. The sandstone is wellsorted, subangular to subrounded, and consists predominantly of quartz. The facies exhibits some fair to good (10 to 15 percent), visible, intergranular porosity and is often oil stained (spotty, patchy, or an even light stain). It is considered to be a good reservoir facies.

In two wells in the Daly field (Townships 8 to 10; Ranges 27 to 29 WPM), unit 1 grades locally near its base to an interbedded sandstone and red, dolomitic, silty shale (Fig. 8). It grades entirely to a conglomerate in 4-29-10-29 WPM, and into an interbedded siltstone and shale north of Daly field in wells at 4-29-15-26 WPM, 8-8-16-27 WPM, 11-9-16-27 WPM and 14-35-16-27 WPM.

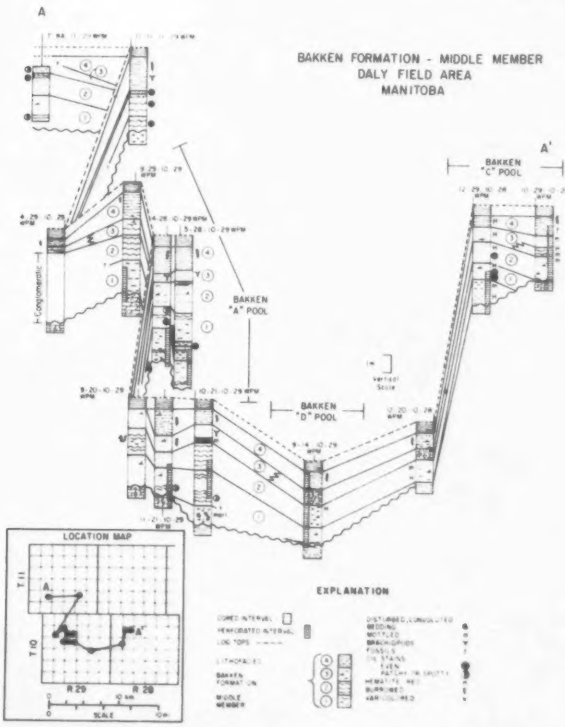


Figure 8: Fence diagram showing the distribution of the lithofacies units of the Middle Member of the Bakken Formation in the Daly field area of Manitoba. Lithofacies represented are described from cores. Sedimentary features, as well as oil staining are indicated.

Unit 1 is present where the Middle Member is thickest. It appears to have been deposited as massive beds in minor erosional or topographic lows on the underlying Devonian surface. Thickness ranges from 0.7 m to 4 m (2 to 13 ft) and averages 2 m (7 ft). The contact of unit 1 with the Three Forks Formation is erosional, as is evidenced by the erosional breccia or conglomerate at the base of unit 1 in several of the cores examined.

Lithofacies Unit 2 is an interbedded sequence of buff, light grey, tan or occasionally "salt and pepper", very fine- to fine-grained, silty sandstone and laminae of medium grey or green, silty shale. It is very thinly, parallel-laminated or thinly, wavy bedded or slightly undulatory. The laminations are 0.5 to 1 cm (0.2 to 0.4 in) in thickness. The sandstone is quartzose, slightly argillaceous, well sorted, angular to subrounded, and generally cemented with dolomite.

Patchy oil staining was noted in many cores. The facies exhibits fair to poor (5 to 10 percent), intergranular porosity. Burrowing is minor. The unit is mottled in places. Convoluted bedding (soft sediment deformation) is also locally present. One well, at 6-7-4-21 WPM, grades to an interbedded grey-green, mottled, dolomitic shale, and mottled, dolomitic, argillaceous siltstone.

Erosional contacts were noted within the unit in core at: 4-29-15-26 WPM and 14-35-16-27 WPM (Fig. 9). The 4-29-15-26 WPM well has a 10 cm (4 in) thick erosional contact near the middle of the unit that consists of brick red, very fine-grained sandstone intraclasts. This contact can be traced to the 14-35-16-27 WPM well, where it is present near the base of the unit. The contact is underlain by a 20 cm (8 in) thick, conglomeratic and brecciated zone characterized by desiccation cracks and sandstone intraclasts that range in size from 3 to 8 cm (1.2 to 3 in). Small-scale ripple cross-laminations were noted at the top of the unit in the well at 5-28-10-29 WPM. Unit 2 is entirely conglomeratic in 4-29-10-29 WPM and comprises green-grey shale intraclasts and silty, dolostone stringers.

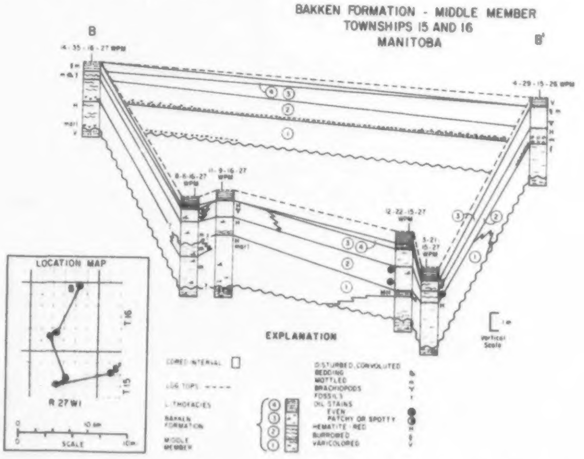


Figure 9: Fence diagram showing the distribution of the lithofacies units of the Middle Member of the Bakken Formation in Townships 15 and 16, Manitoba. Lithofacies represented are described from cores. Sedimentary features, as well, as oil staining are indicated.

Unit 2 is present in all cores examined and ranges in thickness from 0.6 to 1.8 m (2 to 6 ft) averaging 1.2 m (4 ft).

Lithofacies Unit 3 is a buff to light grey, or grey-green, very fine-grained to silty, argillaceous sandstone or siltstone, interbedded with medium grey, silty shale. The sandstone and siltstone occur as discontinuous lenses or faint wavy bedding. Grains are predominantly quartz and generally cemented by dolomite. The shale forms thin laminations or partings.

The unit may contain brachiopods throughout and is commonly burrowed. Pyrite is present in minor amounts. Mottled in places, the unit changes laterally to a silty, argillaceous dolostone or limestone in Township 8, Range 28 WPM and in the Daly field area. Unit 3 is present in all cores examined except 6-21-1-19 WPM and thickness ranges from 0.3 to 2.7 m (1 to 9 ft) and averages 0.8 m (3 ft).

Lithofacies Unit 4 is a massive, light to medium grey, dolomitic, shaly siltstone to silty shale. It is generally burrowed and present in all but seven of the cores examined. Brachiopods occur in distinctive zones or scattered throughout the unit in several cores. Fossil fragments are also a minor component. Thickness ranges from 0.3 to 2 m (1 to 7 ft) and averages 1 m (3 ft). The contact with the overlying shale of the Upper Member is generally sharp.

Bakken Formation: Upper Member

The contact between the Middle and Upper Member is generally sharp. The Upper Member is a black, massive, non-calcareous shale, occasionally fissile or finely banded in places. It may also have a basal, medium grey clay layer that reaches a maximum thickness of 15 cm (6 in). The Upper Member changes in colour from a black shale to a purple, brown, or brownish-red shale in the northern portion of the study area and the eastern half of Daly field. In these areas, the Middle Member is also oxidized, reflecting a period of post-Mississippian/pre-Jurassic subaerial exposure (McCabe, 1959).

SOUTHEASTERN SASKATCHEWAN

Five lithofacies were identified in southeastern Saskatchewan based on the examination and correlation of 35 cores. The subdivisions are based on Christopher (1961), who divided the middle Bakken sandstone member into two units, A and B. Unit B was further subdivided into B1, B2, B3, and B4. For the purpose of this study, Christopher's unit B4 has been renamed as unit C (Table 1). The fence diagram in figure 4 and the structural cross-sections of the Bakken pools (Figs. 10, 11, and 12) correlate these facies on a regional and local basis.

All three members of the Bakken Formation are present in the Saskatchewan portion of the study area; however only the middle Bakken sandstone member was examined in core. The lithofacies of this member are described, in ascending order, as follows:

Bakken Formation: Middle Sandstone Member

Lithofacies Unit A consists generally of a massive, dense, mottled, grey-green siltstone. It is locally argillaceous, highly fossiliferous, very calcareous, with pyrite disseminated throughout. The random orientation of the fossils suggests rapid burial. The basal contact with the lower Bakken shale member is either characterized by scours filled with fossil debris and pebble lags (shale clasts) or by contacts that appear to be gradational. The gradational contacts may be related to the reworking and incorporation of unconsolidated lower Bakken muds into the silts of the middle Bakken sandstone member (Christopher, 1961). Load and flame structures with associated cross-bedding and faint horizontal laminae have been identified in several cores. The upper contact with lithofacies B1 is gradational except in the eastern shelf area. In this area, erosion has progressively truncated unit A and the lower Bakken shale member (Fig. 4). The unit ranges in thickness from 0 to 8.3 m (0 to 27 ft).

Lithofacies Unit B1 consists of parallel interbeds of dark grey shale and buff, or occasionally grey-green, silty sandstone. The disruption of bedding due to very strong bioturbation results in wispy shale "drapes." This disturbed bedding distinguishes this unit from the underlying massive siltstone. The upper and lower contacts are gradational in the western portion of the study area. The upper contact is unconformable where B2 has channelled into B1. The unit is fossiliferous and dolomitic, becoming calcareous with depth. Disseminated pyrite is present. Unit B1 ranges from 1.6 to 14.5 m (5 to 48 ft) in thickness.

Lithofacies Unit B2 is a buff to cream, or greenish-tinted, calcareous sandstone which consists predominantly of quartz with minor feldspar and heavy minerals including disseminated pyrite. It may contain brachiopods and is slightly bioturbated. Numerous diastems are present.

Unit B2 displays a tripartite subdivision. The lower portion is a wavy and flaser bedded, silty sandstone and is transitional with the underlying unit B1. Shale clasts, up to 4 cm (2 in) in diameter, and felsic pebbles are found in channel lags where unit B2 overlies the lower Bakken shale member in the eastern portion of the study area. The middle portion is a coarse-grained, sometimes friable sandstone, and is transitional from the lower portion except where channelling occurs. This sandstone may be massive and/or bedded. Types of bedding include: trough and tabular cross-bedding; horizontal and inclined bedding (up to 25 degrees); and swash cross-stratification. The upper portion is a wavy- and flaser-bedded silty sandstone and is transitional with the middle portion. The upper contact with lithofacies unit B3 is gradational. The B2 unit varies in thickness from 0.4 to 15.6 m (1 to 51 ft).

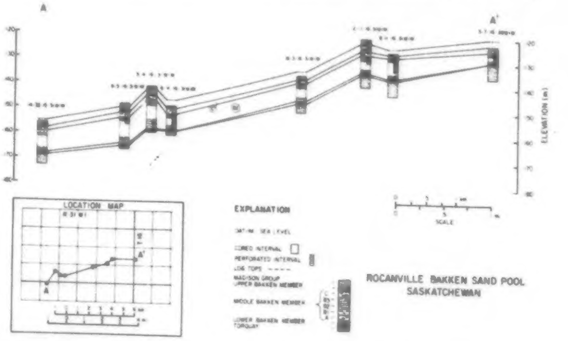


Figure 10: Structural cross-section showing the distribution of lithofacies units across Rocanville field, Saskatchewan. Lithologies represented were described from core.

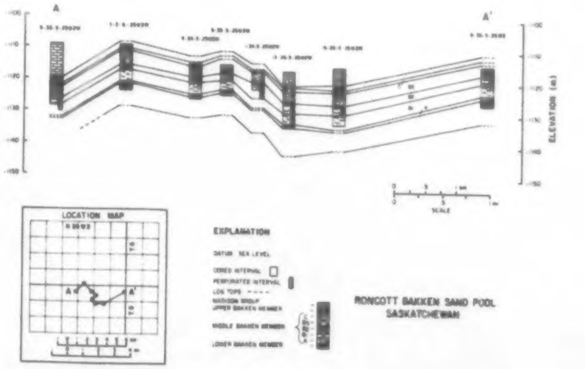


Figure 11: Structural cross-section showing the distribution of lithofacies units across Roncott field, Saskatchewan. Lithologies represented were described from core.

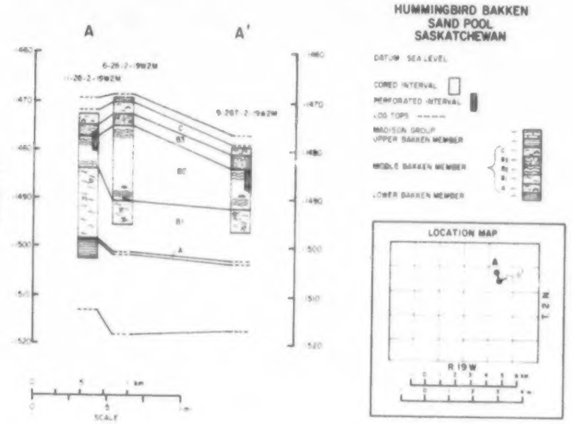


Figure 12: Structural cross-section showing the distribution of lithofacies units across Hummingbird field, Saskatchewan. Lithologies represented were described from core.

Lithofacies Unit B3 consists of parallel interbeds of dark grey shale and buff, slightly bioturbated, silty sandstone. The unit is calcareous. It contains disseminated pyrite and vertical burrows. Load and flame structures occur at the base of the sandy laminae. In the Roncott area (Townships 5 and 6; Range 25 W2M) unit B3 is punctuated by storm-generated density current deposits up to 10 cm thick with what are recognized as the BC(D)E structural divisions of the Bouma sequence. Thickness of B3 ranges from 1 to 4.5 m (3 to 15 ft).

Lithofacies Unit C is a grey-green, massive, dense, mottled, dolomitic, argillaceous siltstone. It is fossiliferous, slightly bioturbated and contains disseminated pyrite. The basal contact is selected either on a massive, mottled siltstone, or where silty rhythmites are encountered. The silty rhythmites range up to 15 cm (6 in) in thickness and are distinguished from B3 rhythmites by their finer grain size and the presence of scattered fossils. One core contained a 6 cm (2 in) thick band of fossil-debris. The upper contact with the upper Bakken shale member is generally sharp and conformable.

DEPOSITIONAL HISTORY OF THE BAKKEN FORMATION - MIDDLE MEMBER

Sandberg *et al.* (1983) developed a depositional model for the rocks of Middle Devonian through Late Mississippian age in the Western Interior. This model is based on eustatic sea level changes. Based on their model, initial sedimentation of the lower Bakken shales resulted from a rapid transgression in conjunction with tectonic activity of the Antler and Acadian orogenic belts. Evidence for this rapid transgression is the sharp disconformity of the Three Forks-Bakken contact along the basin margin. This indicates that the Three Forks was exposed to erosion before shale deposition. Further evidence from cores includes sharp contacts with rip-ups and other erosional features.

The initial phase of black shale deposition appears to have occurred in a shallow sea, based on the presence of benthic fauna in the basal portion of the lower shale member. Paleontological and geochemical data indicate a progressively deepening sea due to on-going transgression. Depth of the water column is still in debate. A stratified water column with anaerobic bottom conditions developed with this deepening event.

The middle member of the Bakken Formation is generally considered to be a transgressive deposit, based on its onlapping relationship with the underlying units (Webster, 1982; 1984; Hester and Schmoker, 1985). However, recent data indicates that the middle member is not entirely transgressive. Thrasher (1985) studied the macrofossils and biostratigraphy of the Bakken. In his study, the middle member was divided into three biostratigraphic units (Table 1). He suggested that unit 1, equivalent to Christopher's unit A, is a regressive deposit. It is the only portion of the middle member that is restricted to the central basin. It does not onlap onto the lower shale and therefore probably represents the low stand of the Bakken seas.

Deposition of the sediments of unit 1 occurred in a moderately well oxygenated, shallow water environment. Thrasher (1985, p. 106) suggested that the random orientation of disarticulated brachiopods probably resulted from "post-mortem current transport followed by rapid burial, and they may represent storm deposits." The environment was probably dysaerobic at times as evidenced by the greenish-grey colouring of the rocks and the presence of pyrite and sulfides in the matrix and on the fossils. Sandberg *et al.* (1983) reported that a long period of continental stability occurred into the Early Mississippian, when seas were at a low stand and the continent was generally flat.

Christopher (1961) interpreted the deposition of the massive siltstone, unit A, to result from a sudden marine transgression. Load structures, pebble and fossil-rich lags at the contact with the underlying shale, combined with randomly oriented fossils, suggest rapid burial as the shoreline advanced eastward, incorporating sediments and dispersing them deeper into the basin. The erosion of unit A in areas more proximal to the basin margin (Fig. 4), makes it difficult to determine the mechanism of deposition in the deeper basin, where unit A is preserved.

In Saskatchewan, subsequent shallowing of the water allowed the development of tidal flats and associated tidal channels, which, when factors such as structure, sedimentology and migration are favorable, form the reservoir. A regressive-transgressive pulse resulted in the deposition of units B1, B2, and B3. Figure 4 shows B2 as a wedge-shaped unit which interfingers with units B1 and B3, until it is no longer recognizable in the deeper parts of the basin. The regression allowed the shoreline to prograde into the basin, depositing the coarse clastic unit B2, in the nearshore environment. The finer-grained sediments of B1 represent the distal equivalent of B2. Subsequent transgression caused the shoreline to move towards the margin of the basin. Unit B3 represents the distal equivalent of B2 during this transgressive phase.

The interbedded laminae of B1 (see Table 1 for North Dakota and Manitoba equivalents) are generally highly bioturbated, suggesting slow sedimentation in a shallow siliciclastic sea (Johnson, 1978). These laminae are transitional to the sands of unit B2 (Table 1), which are typically flaser- to wavy-bedded, reflecting constantly fluctuating, but relatively low-energy conditions in an intertidal flat environment (Elliot, 1978). Elutriation may account for the massive sands of B2. Tidal channels intersecting the tidal flats are more prevalent in the Rocanville area of Saskatchewan, as is swash cross-stratification, which indicates a beach deposit (Harms *et al.*, 1982). In this area, the B2 sands rest unconformably on the lower Bakken shale member or the Torquay Formation.

The B2 sands in the Rocanville (Townships 15 and 16; Range 31 W1M) area of Saskatchewan, which are proximal to the margin of the basin, are very coarse and friable, with relatively good porosity and permeability. In contrast, the porosity and permeability of these sands in the Roncott (Townships 5 and 6; Range 25 W2M and Hummingbird (Township 2; Range 19 W2M) areas, which are much farther from the basin margin, are much less.

Unit B3 (Table 1) is similar in lithology to B1. The units merge in the western and southern portion of the basin (Fig. 4). The B3 unit is slightly to moderately bioturbated with vertical burrows, suggesting a slow rate of sedimentation, in contrast to the slightly faster rate of sedimentation represented by the lower portion of B1, which is intensely bioturbated. The density current deposits interbedded with the deeper water shales of the B3 unit in the Roncott area are probably storm-generated. Walker (1979, p. 94) stated: "The presence of rare "turbidites" would indicate the possibility of density current activity, and would not condemn the entire sequences to deposition in great depths of water."

Unit C (Table 1) is quite thin but widespread throughout the basin (Fig. 4). Currents active during the deposition of this unit were stronger than those occurring during the previous unit. Disarticulated brachiopods present are sorted by size and concentrated into thin, well-sorted beds of grey siltstones and very fine-grained sandstones.

A major eustatic rise in sea level related to further large-scale movement associated with the Antler and Acadian orogenic belt and onlap of the black shale sea followed deposition of the middle member (Sandberg *et al.*, 1983). This sea level rise represents the high stand of the Bakken seas.

The depositional history of the basin is further complicated by the dissolution of salt in the Devonian evaporites. This collapse and associated features can be documented in North Dakota, Manitoba, Saskatchewan, and Montana (McCabe,1959; Christopher, 1961; Anderson and Hunt, 1964; Webster, 1982; 1984; Martiniuk, 1988).

STRUCTURE

A structure map constructed on the Bakken Formation for the study area is shown on figure 13. In North Dakota the Nesson anticline is the most prominent structural feature. This anticline trends north for 176 km (110 mi) from the Killdeer Mountains, Dunn County, to just south of the Canadian border (Fig. 14). Associated with the Nesson anticline is the western Nesson fault, a feature that extends for the length of the fold, and the northwest-trending Antelope anticline. The northeast side of the Antelope structure is also faulted along its length. Episodic movement along the western Nesson fault and the development of the Nesson anticline has been related to basement tectonics (Gerhard *et al.*, 1982; 1987; LeFever *et al.*, 1987).

In the north-central portion of North Dakota, another basement feature that influenced the deposition of the overlying units is the fault that separates the Superior craton from the Trans-Hudson orogenic belt (previously referred to as the Churchill craton or Precambrian province) (Green *et al.*, 1985). Movement of fluids along this fault zone is apparently responsible for the dissolution of salt from the overlying Prairie Formation (Devonian). Evidence for this dissolution can be observed on structure and isopach maps of the formations overlying the Prairie.

Minor southwest-plunging folds are present in Burke County (Townships 161 to 164 N; Ranges 89 to 93 W) (Fig. 15). These folds probably result from basin subsidence, and have had an effect on the depositional patterns of the individual lithofacies, and as well porosity and permeability within the lithofacies.

In Manitoba, structure on the Upper Member is fairly regular and follows the regional Paleozoic tilt, dipping southwest at an average of 6 m/km (32 ft/mile) (Fig. 13). Several features depart from the regional pattern. The Virden (Townships 9 to 11; Ranges 25 to 27 WPM) and Daly (Townships 8 to 10; Ranges 27 to 29 WPM) fields are separated by a prominent north-south trending syncline which lies along the Birdtail-Waskada axis. The latter is a north-trending zone, approximately 32 km (20 mi) east of the Saskatchewan border, along which numerous local structure and isopach anomalies in the Devonian and later strata are concentrated. These anomalies may be a direct or indirect result of salt dissolution resulting from minor tectonic movement along the boundary zone between the Churchill and Superior provinces (Trans-Hudson orogenic belt) (McCabe, 1966; Green *et al.*, 1985). The axis coincides with the crustal boundary and the present solution edge of the Prairie Formation salt section (McCabe, 1966). The anomalies have been important in controlling the accumulation of oil in the Daly, Virden, and Waskada (Townships 1 and 2; Ranges 25 to 27 WPM) producing areas of southwestern Manitoba (McCabe, 1963).

Within the Daly field itself, several minor structural highs are coincident with areas of Bakken production. Within the Waskada field area a minor south-southwest plunging structural nose is coincident with anomalous Bakken and Three Forks thickening.

The Saskatchewan portion of the study area is divided by the Elbow-Hummingbird monoclinal flexure into two structural provinces (Christopher, 1961). These provinces are defined by the presence or absence of Middle Devonian evaporite beds (Fig. 14). The evaporites terminate along a northwesterly trend which approximately coincides with the Elbow-Hummingbird flexure. The eastern side forms an easterly striking monocline with associated subsidiary anticlinal and synclinal structures (including the syncline that underlies the Rocanville field). Devonian evaporites underlie the eastern portion. The western side has an irregular surface with easterly and northeasterly trends. The evaporites are absent from the western side. The Elbow-Hummingbird monoclinal flexure forms the eastern limb of the Hummingbird synclinorium in the southernmost portion of the study area.

Adjacent to the Hummingbird synclinorium is the Roncott anticlinorium. This is a broad, 48 km (30 mi) wide fold plunging in a south-southeasterly direction. The structure exists due to the presence of an outlier of Devonian evaporites. Overlying sediments are draped over the outlier, forming the feature.

Other notable features include the Rocanville-Torquay trend, the Regina- Melville platform, and the Herald and Torquay embayments (Christopher, 1961). The Rocanville-Torquay trend consists of a northeast-trending series of depressions outlined by producing wells (Fig. 14). The Regina-Melville platform represents "a broad undifferentiated sedimentary area between more basinal localities to the north and south" (Christopher, 1961.

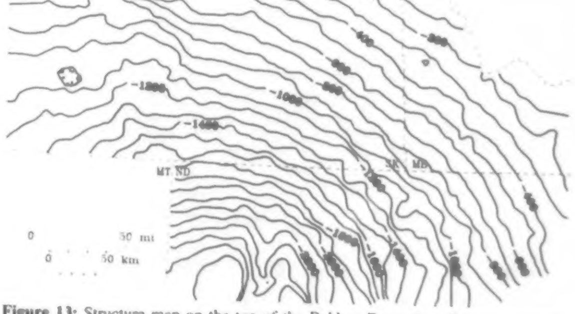


Figure 13: Structure map on the top of the Bakken Formation. Contour interval is 100 m (328 ft). Dashed line in northeastern Manitoba is the erosional limit of the Bakken Formation.

p. 59). The Herald and Torquay embayments represent the expression of the Williston Basin in southern Saskatchewan.

ISOPACHS

Distribution of the middle member of the Bakken Formation throughout the study area is shown in figure 16. In North Dakota the Bakken isopachs show depositional patterns related to the Elk Point Basin sedimentation. The depocenter of the middle member is elongated north-south and is located just to the east of the Nesson anticline. This is similar to isopach maps of the lower member. However, these trends are not apparent in the upper member, which has a poorly defined depocenter. The middle member, from its maximum thickness of 27 m (87 ft), thins gradually away from the depocenter towards its depositional limit.

Abrupt thickening and thinning of the middle member in the north-central portion of North Dakota are probably related to the dissolution of the underlying Devonian Prairie salt and subsequent collapse. Isopach maps of all three members, as well as underlying and overlying formations, display abrupt thickness changes. This is further supported by isolated occurrences of the lower member throughout north-central North Dakota.

Examination of an isopach map constructed for the middle member sandstone, unit 3, shows thinning over pre-existing structural features, such as the Nesson anticline and a series of folds in Burke County (Fig. 7). This is supported by the examination of available cores. The isopach of unit 3 also shows that its distribution is not as extensive as the distribution of the entire middle member which continues south of the study area.

In Manitoba, the isopach trends of the Bakken Formation are generally northwest-southeast, reflecting deposition related to the Elk Point Basin (Fig. 2). The most significant changes in thickness of the Bakken occur within the middle member, which

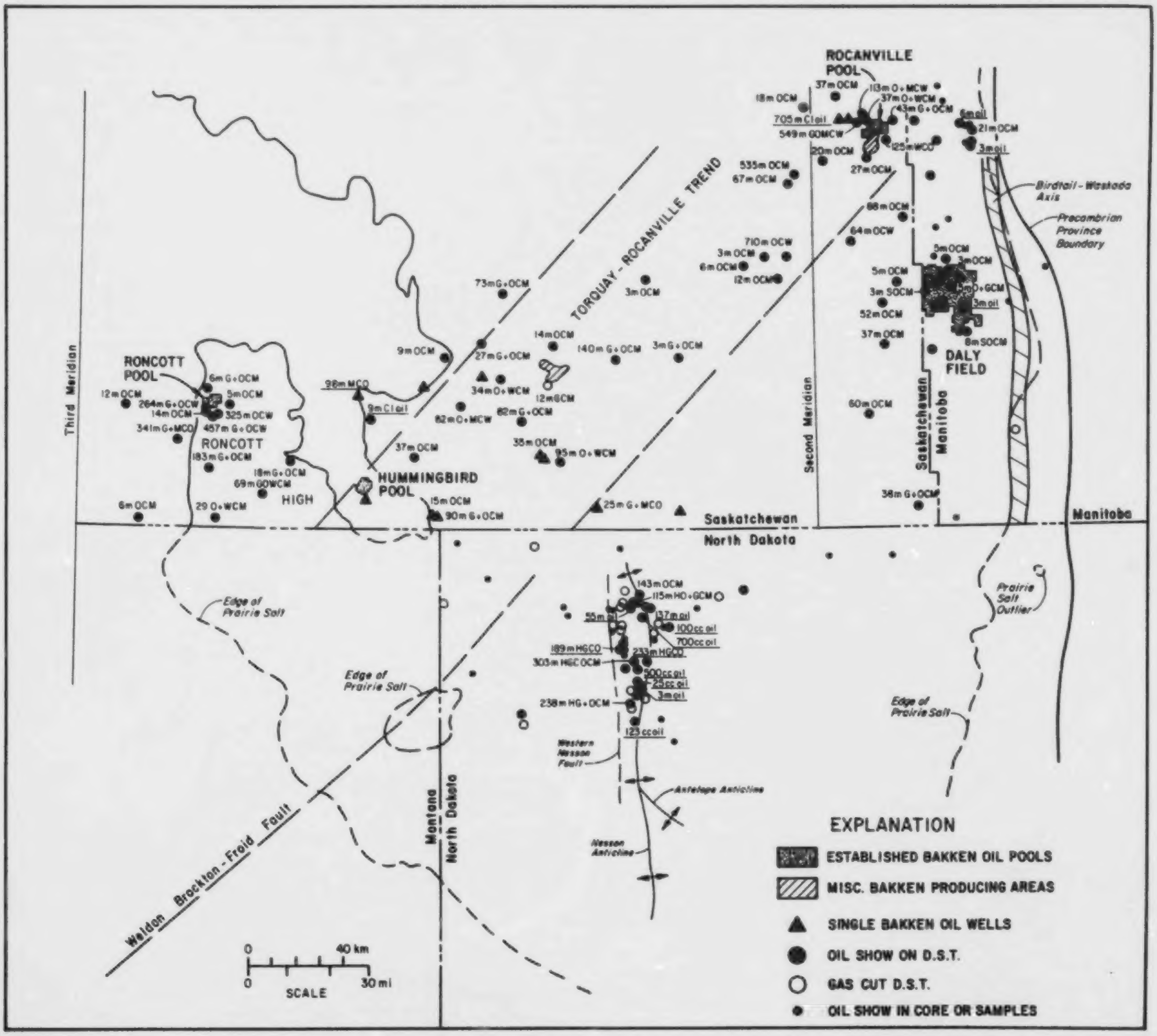


Figure 14: Diagram showing the relationship of structural features to oil shows and oil production within the study area. Major oilfields producing from the middle member of the Bakken Formation, oil shows from drill stem tests, cores or samples, and gas shows are indicated. Limit of the Prairie Formation salt is from Holter (1969), Anderson and Swinehart (1979), and McCabe (1980).

represents a major portion of total Bakken thickness (Fig. 16). The middle member averages 4 m (13 ft) in thickness; the upper member 2 m (7 ft). Sedimentation of the Bakken is controlled, in part, by the paleotopography of the underlying Devonian erosion surface. Minor thickening of the middle member may represent the infill of clastics into these erosional or paleotopographic lows.

Local variations in thickness in both the Bakken and Three Forks formations are evident throughout the study area in Manitoba. The greatest thickness variations occur in the Waskada field area where as much as 26 m (85 ft) has been added to the total Bakken and over 79 m (259 ft) added to the Three Forks. These anomalies are believed to be the result of the multiple-stage dissolution of the Devonian Prairie Formation salt section during the deposition of the Bakken and Three Forks formations. The local occurrence of the lower member in the Waskada area may also be accounted for by Prairie salt dissolution during early Bakken sedimentation (Martiniuk, 1988).

In Saskatchewan the isopach lines trend northwesterly (Figs. 2 and 16). In the eastern portion of the study area thicknesses gradually thin to the northeast. This changes abruptly west of the Elbow-Hummingbird monoclinal flexure. There are a series of thicks and thins that are possibly related to the dissolution of the Devonian Prairie Evaporite.

The Torquay-Rocanville trend is readily apparent on the isopach of the Bakken Formation (Fig. 2). This feature trends to the northeast from the Montana and North Dakota state lines. It is also present but not as well developed on the isopach of the middle member (Fig. 16).

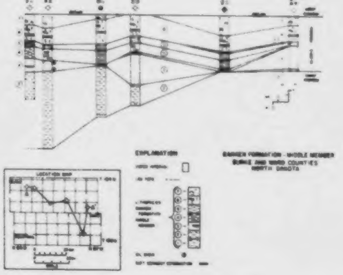


Figure 15: Fence diagram showing the distribution of lithofacies units across Burke and Ward counties, North Dakota. Lithologies represented are described from core.

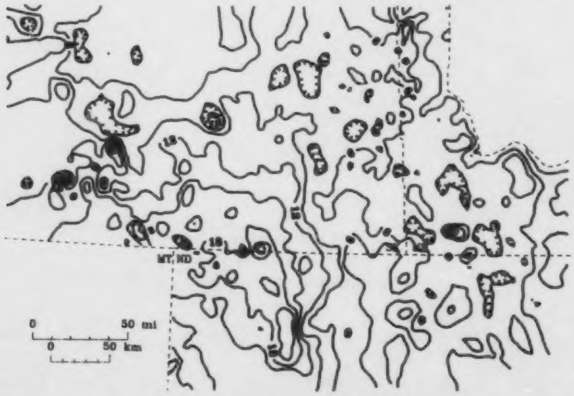


Figure 16: Isopach map of the middle member of the Bakken Formation for the study area. Contour interval is 3 m (10 ft). Dashed line in northeastern Manitoba is the erosional limit of the Bakken Formation.

PRODUCTION AND RESERVOIR CHARACTERISTICS

NORTH DAKOTA

The middle member of the Bakken Formation produces or has produced in nine fields situated along the northern portion of the Nesson anticline (Fig. 17). Fourteen wells scattered throughout these fields have a cumulative production through December 1990 of 85,334 m^3 (536,753 bbls).

Until recently, the Bakken Formation in North Dakota was generally considered not to be an economic venture as a primary target. Wells with associated Bakken production generally have multiple completions. Recompletions in the Bakken are common as wells reach their economic limit. Only one well (NESE Sec. 8, T.159N., R.95W.) in the study area completed in the middle member had the Bakken as its primary objective.

Three cycles of Bakken development can be seen in an examination of productive wells. The initial completion in the middle member occurred in 1963. Considered to be uneconomic, the well, a Duperow test, produced 118 m^3 (744 bbls) from the middle member before recompletion in the Madison. No further attempts occurred until the late 1970's when the price of oil increased significantly. This increased the profitability on marginal wells and encouraged exploration in the Bakken. The decline in oil prices resulted in another break in Bakken drilling. Attention has been redirected toward the Bakken once again, since the advent of the Bakken shale horizontal play.

Stoneview Field

Stoneview field (Townships 160 and 161 N; Ranges 94 and 95 W) is located along the northern tip of the Nesson anticline (Fig. 17). It is the most prolific of the Bakken pools, producing from the middle member in North Dakota with a cumulative production of 33,955 m^3 (213,580 bbls) of oil.

Three wells currently produce along the northern and western boundaries of the field. Perforated intervals include the upper shale to the base of the sandstone unit (unit 3). The best producer (SESW Sec. 30, T.161N., R.94W.) in the pool was perforated only in the sandstone unit and has attained a cumulative production of 11,454 m^3 (72,049 bbls).

All three wells were acidized during completion. Acid amounts range from 1,000 to 3,000 gallons of 15 percent HCl. One well (NENW Sec. 31, T.161N., R.94W.) also underwent a sand/oil fracture treatment (80,625 lbs of propant in 33,348 gallons of fluid). The best producer in the pool underwent the least amount of stimulation (1,000 gallons of acid). This may be explained by the restriction of the perforation to the sandstone unit. The other two wells also included perforations in the upper shale. Acidization of the upper shale results in the release of fines which may partially plug effective porosity.

Reservoir characteristics for Stoneview field are summarized in Table 2.

West Tioga

Production in West Tioga is from three wells situated in the central portion of the field (Townships 157 and 158 N; Range 95 W). The cumulative production from 1976 through December 1990 was 16,403 m^3 (103,175 bbls).

Reservoir characteristics for this field are limited. The perforated interval is restricted to the middle member sandstone, unit 3, with an average net pay of 3 m (10 ft). Gravity of the oil was reported as 42.0° API.

Only one well (NENE Sec. 9, T.157N., R.95W.) has a lengthy production history. Completed in February 1976, the well has a cumulative production of 11,669 m^3 (73,400 bbls). This well was stimulated with two acid treatments (4,000 gallons of 15 percent HCl) and a gelled water/sand fracture treatment. The next well was completed in 1987 and was acidized with 500 gallons of acid.

Other Fields

Production in the remaining fields is restricted to one or two wells. These wells are responsible for the remaining 34,976 m^3 (219,998 bbls) of oil produced from the middle member of the Bakken Formation.

The Bakken pool in North Tioga (Townships 159 and 160 N; Ranges 94 and 95 W) consists of two wells with a cumulative production of 6,929 m^3 (43,585 bbls). The earliest well, completed in 1987, perforated the entire Bakken interval. Perforations in the second well, completed in 1988, are restricted to unit 3. Both wells were stimulated with a sand-oil fracture treatment.

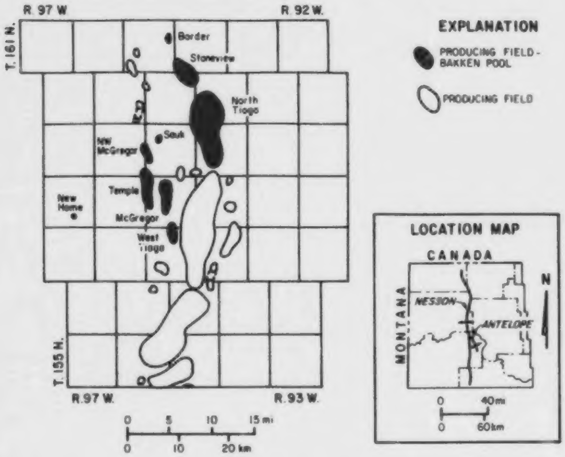


Figure 17: Map showing the distribution of producing oil fields along the northern Nesson anticline. Fields producing from the middle member of the Bakken Formation are indicated in black.

Table 2: Reservoir properties for North Dakota oil fields producing from the sandstone, unit 3, within the middle member of the Bakken Formation.

	Temple	North Tioga	Stoneview
Spacing Unit	160	320	320
No. of Producing Wells	1	2	3
Average Net Pay (ft)	12	12.5	6
Average Porosity (%)	7	2.3	12
Average Water Saturation (%)(S_w)	80	15	21
Oil Gravity, API @ 60°	40	39.7	44.0
Reservoir Drive Mechanism	Solution Gas	Solution Gas/ Fluid Drive	Solution Gas
Volumetrics, OOIP, bbls	139,000	472,798	1,765,000*
Recoverable Oil	14,000	47,279	264,765*
Recovery Factor (%)	10	10	15
Cumulative Production Oil (12/1990) Water (12/1990)	 4,862 1,704	 43,585 5,770	 213,580 12,198

*Volumetrics are calculated for a spacing unit of 160 acres, instead of the assigned 320 acre spacing.

Temple field Bakken pool (Township 158 N; Ranges 95 and 96 W) consists of two wells which were drilled in 1989. The initial recompletion was unsuccessful with low oil-high water production and was plugged and abandoned eight months later. This well was perforated in the lower Lodgepole Formation, not the Bakken. The second well was perforated in the upper shale to the base of the sandstone, unit 3. Reservoir data for this well are presented in Table 2.

The Bakken pools in the remaining fields consist of one well each. Perforated intervals cover the upper shale to the base of the sandstone or at least the middle member. All of the wells were stimulated with an acid and/or fracture treatment. Limited production was yielded by most of the pools with the exception of the wells in Sauk (Townships 159 N; Range 95 W) and McGregor (Townships 158 and 159 N; Range 95 W) fields.

SOUTHWESTERN MANITOBA

Daly Field

Production was first established in southwestern Manitoba in the Daly field in 1985, following the Newscope Resources Limited discovery at 13-21-10-29 WPM. The well was completed in the Middle Member with 2 m (7 ft) of net pay. Prior to the Bakken discovery, the Daly field was productive only in the Lodgepole Formation. Bakken and Lodgepole production is presently commingled in several wells.

Bakken oil in southwestern Manitoba is 824 g/m^3 (40.2° A.P.I.) and has a sulphur content of 1.30 g/kg. The shallow depth to production and high quality of oil make the prospect for Bakken in Manitoba attractive. Average depth to the producing zones of the Middle Member is 874 m (2,867 ft).

Seven Bakken pools have been established in the Daly field since the initial discovery: A, B, C, D, E, G and H (Fig. 18). In total, 35 wells were active Bakken producers as of January, 1991. As of December 31, 1990, the Bakken had produced 83,700 m^3 (526,473 bbls) of oil. Figures 19 and 20 summarize cumulative and average daily Bakken production for the Daly field.

Proven remaining established reserves for the Bakken in the Daly field as of December 31, 1990, were 43,200 m^3 (271,728 bbls)(1990 Manitoba Energy and Mines reserves estimates). Reservoir properties described for the three multi-well Bakken pools in the Daly field (A, B and D) are given in Table 3.

Bakken oil reservoirs within the Daly field appear to be controlled primarily by stratigraphic factors. They are localized within paleotopographic lows of the underlying Devonian erosional surface, where the sandstones are thickest.

Unit 1 is the primary reservoir facies of the middle member and the one in which there are the majority of completions. Based on core analyses, porosities range from approximately 15 - 20 percent within the unit. Oil shows have also been noted in unit 2, although, permeabilities are generally lower within this facies due to more silt and finely argillaceous laminae.

The distribution of unit 1 is related to the paleotopography of the underlying Three Forks Formation. It is present and a potential reservoir in the Daly Field where it has infilled minor erosional lows on the underlying Devonian erosional surface.

Stratigraphic trapping within the middle member in Daly field is created by the lateral pinchout and variation of unit 1 in combination with an upward fining to the very fine-grained, finely laminated sandstone of unit 2.

The fence diagram (Fig. 8) shows the productive facies within the middle member in the Daly Field. In the Bakken A Pool it is difficult to distinguish the lithology of unit 1 from that of the underlying Three Forks Formation. The lower contact is questionable in the wells at 10-21-10-29 WPM, 4-28-10-29 WPM and 5-28-10-29 WPM. However, unit 1 is interpreted in these wells to have changed laterally in its lower portion, into an interbedded, siltstone and blocky, green shale or marl. This has limited the eastern extent of the reservoir by creating vertical and lateral permeability barriers.

The western limit of the Bakken A pool is defined by the well at 4-29-10-29 WPM, where both units 1 and 2 are completely conglomeratic. The northwestern limit of the pool is undefined and offers potential for future development.

The Bakken D Pool, as represented by the well at 9-14-10-29 WPM, shows several lateral changes in units 1, 2 and 3 from the wells at the Bakken A Pool. Unit 1 has changed from a marly, interbedded, very fine-grained sandstone and shale to a very fine- to fine-grained, banded sandstone. Unit 2 has changed to a very fine-grained, mottled, hematite-red sandstone with green, shale laminations near the base, and unit 3 lithology from a very fine-grained, mottled sandstone to an argillaceous, silty dolostone.

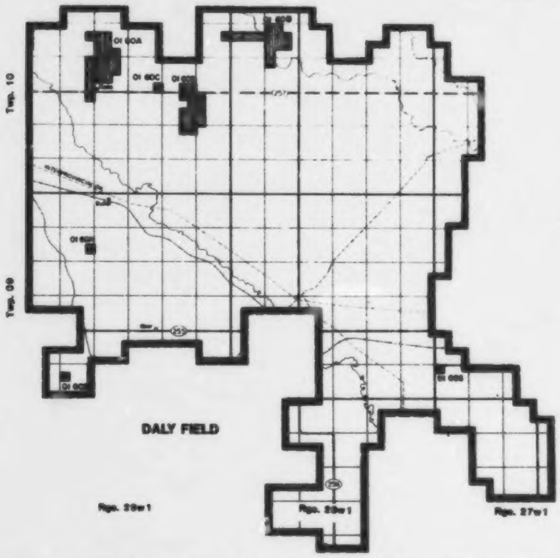


Figure 18: Map showing the producing wells as of January 1991 in Daly field, Manitoba (Martiniuk, 1988).

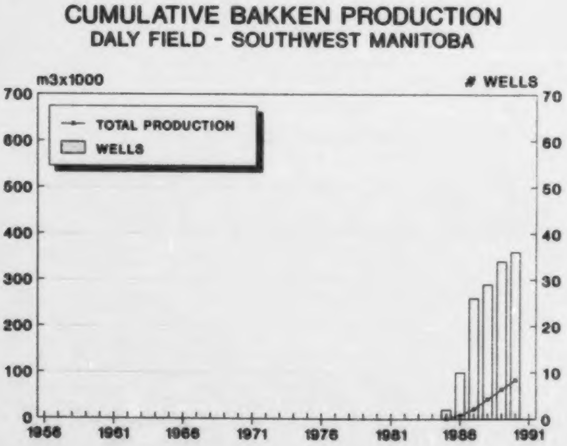


Figure 19: Cumulative production from the Bakken in Daly field in southwestern Manitoba.

The lithology of the Bakken B Pool is similar to the Bakken A Pool; however, all or most of the units in the middle member are hematite-red, which indicate that these units have undergone oxidation or exposure that appears to have occurred less extensively elsewhere in the Daly field area.

Structural factors are of secondary importance in the accumulation of Bakken oil in the Daly field. Minor structural closure is evident on highs on the upper member at the Bakken B Pool. Minor structural highs or noses on the upper member are also noted at the Bakken A, C and D Pools (Fig. 13).

It should be noted that a potential reservoir also exists below the Devonian erosional surface within a vuggy dolostone. The unit was perforated along with unit 1 in the productive Newscope Opinac Daly 11-21-10-29 WPM well.

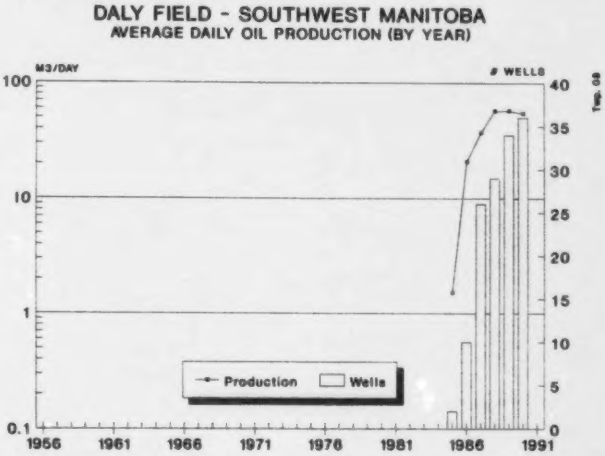


Figure 20: Average daily oil production by year for Daly field in southwestern Manitoba.

Table 3: Reservoir properties for the Bakken pools in Daly field, Manitoba.

	Daly Bakken A Pool	Daly Bakken B Pool	Daly Bakken D Pool
Area (ha)	336	240	240
No. of Producing Wells	15	7	12
Average Net Pay (m)	1.6	1.8	2.0
Average Porosity (%)	17	16	15
Average Water Saturation (%)	45	54	46
Shrinkage Factor (fr.)	0.89	0.89	0.89
Original Oil-in-Place ($10^3 m^3$)	1125.2	492.0	336.4
Recoverable Reserves ($10^3 m^3$)	62.8	11.7	51.4*
Primary Recovery Factor (%) (solution gas drive)	5.6	2.4	15.3
Cumulative Oil Production ($10^3 m^3$)	46.6	8.3	27.8
Remaining Recoverable Reserves ($10^3 m^3$)	16.2	3.4	23.6

* A pilot waterflood program has been recently initiated in the Bakken D Pool. Injection commenced on an inverted 7-spot injection pattern in July, 1990. Waterflood response was observed in early 1991. Between July, 1990 and February, 1991, daily production increased 16% from 15.4 m^3/d to 17.8 m^3/d. Cumulative voidage-replacement to March, 1991 is 0.22. Preliminary estimates for waterflood recovery are 27.5%, compared to a primary recovery for the pool of 15.3%.

SOUTHEASTERN SASKATCHEWAN

Cumulative Bakken oil production from southeast Saskatchewan totalled 601,232 m³ (3.8 million bbls) at the end of 1989. Figure 21 indicates a sharp increase in yearly Bakken production beginning in 1985. A portion of this increase can be attributed to an infill program at Rocanville. However, the majority of the increase is due to the development of a new Bakken pool at Hummingbird as well as scattered Bakken discoveries throughout the area (Fig. 14). Economic production from the Bakken throughout southeastern Saskatchewan can now be found outside the traditional Bakken producing areas of Roncott and Rocanville

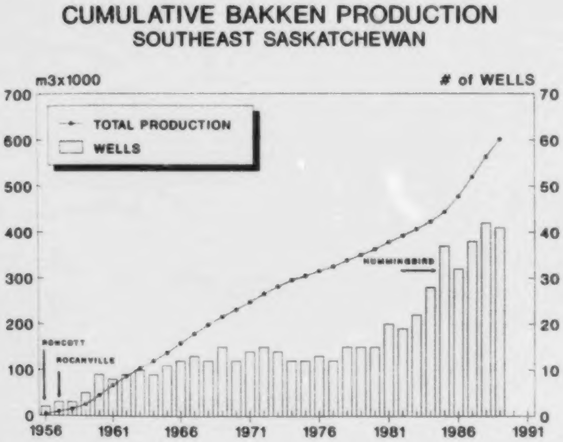


Figure 21: Cumulative production from the Bakken Formation in southeastern Saskatchewan.

The majority of production originates in the medial B2 sandstone unit where permeabilities and porosities are the greatest. However, oil staining has been noticed, and production has been attained, from silty intervals within the B1 and B3 units. This is possible due to the interfingering nature of these three units (Christopher, 1961). The future potential for Bakken production along this trend is enhanced by the occurrence of these stratigraphic traps and provides new possibilities for exploration beyond the areas of known structural traps.

Rocanville Pool

Rocanville is the most prolific Bakken reservoir within southeastern Saskatchewan (Fig. 22). Cumulative production as of October, 1990 reached 339,165 m³ (2.1 million bbls). Remaining reserves are 102,000 m³ (641,580 bbls) from initial reserves of 441,000 m³ (2.8 million bbls) or 19 percent of original oil-in-place. The original oil-in-place as calculated by Saskatchewan Energy and Mines (1989) is 2.326 million m³ (14.6 million bbls).

The Rocanville reservoir is located entirely within the B2 unit of the middle Bakken sandstone member where it exhibits typical channel cross-bedding, flaser bedding and/or rhythmites (Fig. 10). This lithology accounts for the excellent permeabilities in this reservoir which range from 30 md to as high as 1000 md with an average of 65 md. Permeability may have been enhanced by dolomitization of the original calcium carbonate cement (Kent, 1984). The average net pay is 5.6 m (18 ft). The average porosity is 20 percent (Saskatchewan Energy Mines, 1989).

Pressure history on this pool indicates that the reservoir oil was highly undersaturated at original conditions and that the pool is subject only to rock and liquid expansion along with a partial natural water influx drive mechanism (Gillard and Jordan, 1984). Excellent reservoir permeability would therefore account for the moderate oil recovery.

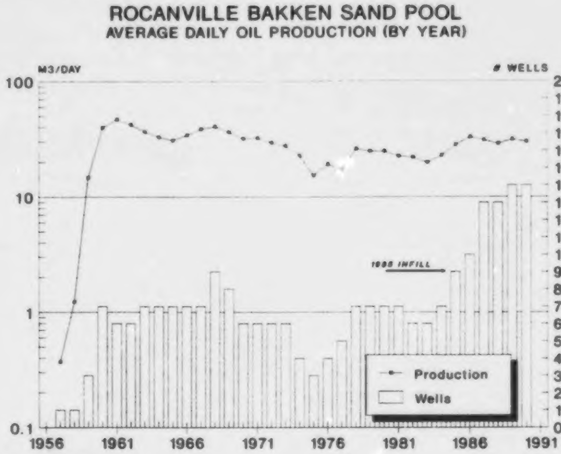


Figure 22: Average daily oil production by year for the Rocanville Bakken sand pool in Saskatchewan.

Roncott Pool

Roncott is the oldest producing Bakken reservoir in southeastern Saskatchewan, and has produced since 1956 (Fig. 23). It has a cumulative production of 104,843 m³ (659,462 bbls). The pool may be near the end of its productive life with remaining reserves of only 11,000 m³ (69,190 bbls) from initial reserves of 116,000 m³ (729,640 bbls). Original oil-in-place is calculated at 1.037 million m³ (6.5 million bbls) (Saskatchewan Energy and Mines, 1989) with estimated recoveries of 11 percent. The virtues of infill drilling should be reviewed to determine whether or not the capture of wedge and/or unswept oil is an economic option.

The majority of the wells in the Roncott reservoir are producing from the B2 unit, however, the perforated intervals of some wells have crossed over into either the B1 unit and/or the B3 unit (Fig. 11). Examination of core from these wells shows oil staining in coarser-grained beds within these units. Permeabilities range from 0.75 md to 75 md, averaging 6 md. These permeabilities are extremely low compared to an average of

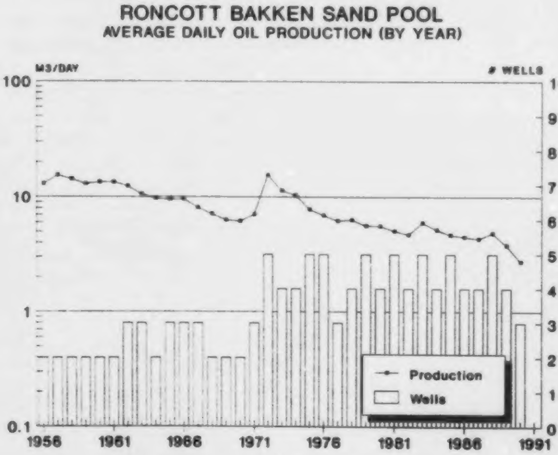


Figure 23: Average daily oil production by year for the Roncott Bakken sand pool in Saskatchewan.

65 md in the Rocanville reservoir. The difference reflects the interfingering of units B1, B2, and B3 along the distal margins of the clastic wedge, while the proximal reservoir at Rocanville is composed entirely of B2. Net pay averages 4.5 m (15 ft) and porosities average 14 percent. Porosity enhancement by dolomitization is more evident here than at Rocanville.

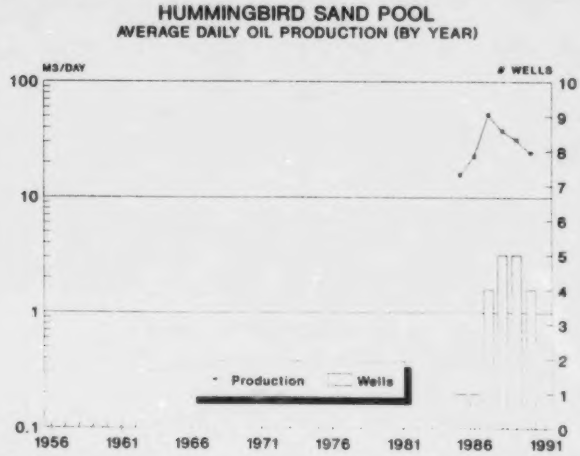


Figure 24: Average daily oil production by year for the Hummingbird Bakken sand pool in Saskatchewan.

Hummingbird Pool

The newest pool of established Bakken production is on the Hummingbird structure, where Birdbear and Ratcliffe production already exist. Bakken production at Hummingbird began in 1985 and rose dramatically over the next two years. Since 1987 a rather steep production decline has persisted (Fig. 24).

Cumulative production is 60,210 m^3 (378,720 bbls) to October, 1990. This represents well over half of the initial reserves of 98,000 m^3 (616,420 bbls) and leaves only 38,000 m^3 (239,020 bbls) remaining to be produced. Saskatchewan Energy and Mines (1989) has calculated original oil-in-place at 525,000 m^3 (3.3 million bbls) which places the recovery factor at 19 percent.

The majority of production is from the B2 unit (Fig. 12). There are also shows in the lowest portion of the B3 unit, where at least two wells have been completed. A distal equivalent of the producing channel sands at Rocanville, the Hummingbird reservoir has a higher argillaceous content. The sedimentary structures that dominate the Hummingbird reservoir include flaser-type bedding and wavy laminated sands and silts instead of the channel cross-bedding present at Rocanville. Porosities and permeabilities at Hummingbird are lower than the Rocanville reservoir. The average porosity and permeability at Hummingbird is 9 percent (Saskatchewan Energy and Mines, 1989) and 5 md, respectively. Average net pay is 4.9 m (16 ft).

Miscellaneous Production

Miscellaneous single well production and undesignated small pool production accounted for 114,335 m^3 (719,167 bbls) of the total Bakken production in southeastern Saskatchewan as of December, 1989. The most significant production is from a small cluster of 12 wells in the North Midale area (Townships 6 and 7, Range 11 W2M) that produced 56,599 m^3 (356,007 bbls).

The majority of this production is found within the Rocanville-Torquay structural trend (Fig. 14) (Christopher, 1961). The few exceptions to this rule are found on trend with the Brockton-Froid Fault zone and the Weldon Fault of northeastern Montana suggesting that these features may extend into southeastern Saskatchewan.

POTENTIAL

North Dakota

Potential exists for production from unit 3, within the middle member of the Bakken Formation in North Dakota. Core examination shows that the depositional pattern of the lower lithofacies, units 1 through 5, were affected by pre-existing structure and topography. Thinning of the sandstone, unit 3 over structural noses is evidenced by cores examined from Burke County (Fig. 15). This relationship strongly suggests a potential for updip stratigraphic pinch-out of the sandstone unit against impermeable rocks. When this situation is combined with a source rock, the Bakken shales, a potential for production is present. The sandstone unit is in contact with the upper Bakken shales of the study area. Fractures present locally also form other migration pathways. These factors can also be applied to other potentially productive lithofacies of the middle member.

A favorable property of the sandstone unit is its response to stimulation. The sandstone unit reacts well to stimulation by acid and/or sand-oil fracture treatments that can greatly increase the effective drainage area and enhance production.

Southwestern Manitoba

The Daly field has been developing continuously since Bakken oil was discovered in 1985. The area has been attractive due to the high quality of Bakken oil and the prospect for uphole completion in shallower zones within the overlying Lodgepole Formation.

Most of the Bakken pools presently established in the Daly field have been defined. Potential for future Bakken development remains however, in the northwestward extension of the Bakken A Pool into Township 11, Range 29 WPM (Fig. 8) where unit 1 is present.

Eastward, in the Virden field (Townships 9 to 11; Ranges 25 to 27 WPM), the Bakken remains largely untested. This is an area of Bakken thickening and of known salt collapse features within the shallower Lodgepole Formation. Possibility for a Bakken subcrop play may exist along the eastern edge of the field (Fig. 16).

Northward, toward the northeastern limit of the Bakken Formation, several free-oil recoveries have been made from drill stem tests in the middle member; specifically, from wells in Townships 15 and 16, Range 27 WPM (Fig. 14). The locality features a Bakken structural nose (Fig. 13). Minor erosional breaks within units 1 and 2 and lateral lithologic changes from fine grained sandstone to argillaceous, laminated siltstone, offer potential for a stratigraphic trapping eastward of this area. Several completions have been attempted but have not been successful due to high water recoveries.

Southeastern Saskatchewan

The potential for exploration and new discoveries in the Bakken is immense. Oil shows throughout the Saskatchewan portion of the Williston Basin as well as scattered single well production along the Bakken production trend (Fig. 14) suggest that significant amounts of undiscovered Bakken oil remain to be found.

Bakken sourced oils are characterized by very long migration pathways (Osadetz *et al.*, this volume), having been generated in the deepest portion of the basin, south of the Missouri River, and migrated radially as far away as Rocanville, over 200 km (124 mi). The length of migration may be a factor in determining API gravity of the oil. Rocanville oil has a gravity of 36° API whereas Roncott and Hummingbird produces 40° API oil. Even closer to the source, the previously mentioned North Midale wells produce 41° API oil and are situated directly north of the Bakken hydrocarbon generation region as described by Osadetz *et al.* (in review) (Fig. 25). Wells tested at Torquay and north of the international border at Roche Percee had 43° and 45° API oil, respectively.

Structural features that aid and control entrapment of Bakken oil include the linear depressions associated with the Rocanville-Torquay structural trend (Christopher, 1961), as well as the reversal of regional dip caused by flexure of the Bakken on the flanks of the Roncott high. These structures probably did not aid in the migration of oil from the center of the basin, but do provide favorable conditions for structural entrapment.

The interfingering of the medial B2 sandstone wedge along with the laminated shales and silts of the B1 and B3 units, provides another avenue of entrapment in areas where structural features may not exist.

SOURCE ROCK AND MIGRATION PATHS

The Bakken Formation has been documented as an excellent petroleum source rock within the Williston Basin (Dow, 1974; Williams, 1974; Meissner, 1978; Webster, 1984; Hester and Schmoker, 1985). The black shales are the richest and most widespread rocks in the Williston Basin (Osadetz and Snowdon, in review). The upper and lower shales of the Bakken Formation yield rich Type II sources (Dow, 1974; Webster, 1984; Osadetz and Snowdon, in review). The upper shale is generally richer than that of the lower shale.

Osadetz and Snowdon (*op. cit.*) concluded that thermal maturity of source rocks in the Williston Basin was strongly affected by elevated heat flows, notably along the Nesson crustal structure in North Dakota. Enhanced maturation resulted in Type II oil windows occurring at much shallower than expected depths, at 2,300 m (7,544 ft). The Nesson anticline is believed to have had an important control on hydrocarbon potential.

Production from the Bakken Formation occur at long distances from the regions of significant hydrocarbon generation (vitrinite reflectance = 0.7 percent) and expulsion threshold (vitrinite reflectance = 0.9 percent) (Osadetz *et al.*, in review). Osadetz *et al.* (in review, p. 84) noted that the "region sufficiently mature to expel hydrocarbons occurs in an area south of the Missouri River, a region unfavourably disposed to migrate oils into northeastern Williston Basin. Enhanced maturities are associated with high heat flows in the region of the North American Central Plains conductivity anomaly" (Fig. 25). This anomaly is a 2,000 km (3,220 mi) long, 80 km (129 mi) wide feature along which coincidental electrical conductivity and heat flow anomalies occur. The Nesson structure is associated with this feature. In Canada, Osadetz *et al.* (in review, p. 31) noted that ".... Family B oils occur exclusively in the Bakken Formation reservoirs at Roncott, Daly and Rocanville fields". Production from the Hummingbird pool are Family C oils. Osadetz *et al.* (in review) proposed a long distance migration path for oil to the Roncott, Rocanville and Daly fields (Fig. 25).

It is possible that until 1 (Manitoba) and its correlatives in Saskatchewan and North Dakota, as well as channel sands noted in these areas, provided the necessary conduit for oil migration and may represent the "lowstand system" proposed by Osadetz *et al.* (in review) (Fig. 25) These facies, where developed in erosional and paleotopographic lows on the underlying Devonian surface, are coincident with productive areas in Saskatchewan and Manitoba.

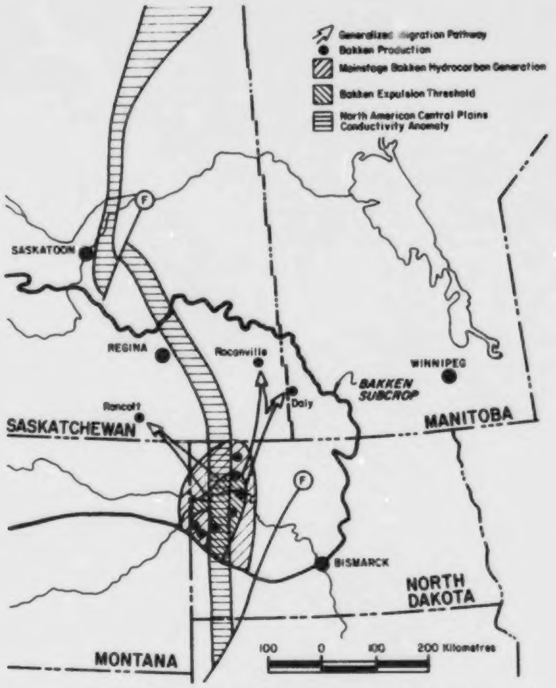


Figure 25: Diagram showing the area of mainstage hydrocarbon generation in the Bakken Formation. The area where the Bakken has exceeded its expulsion threshold is also indicated, as well as, the generalized migration pathway (from Osadetz *et al.*, 1991).

SUMMARY

1. The Bakken Formation consists of three members, a lower and an upper shale and a middle member. The middle member consists of multiple lithofacies which are correlative across the study area.
2. The middle member, overall, is considered to be a transgressive deposit, based on its onlapping relationship with the lower member. The lowermost unit of the middle member is probably regressive. It does not onlap the lower shale and therefore, probably represents the low stand of Bakken seas.
3. Smaller scale transgressive and regressive pulses are exhibited by the deposition of the upper portions of the middle member. The sandstone units represent intertidal deposits with associated tidal channels. Storm deposits are also present, locally.
4. The deposition of the middle member is further complicated by the dissolution of the Devonian evaporites.
5. Topography and structure of the underlying Devonian formations show a strong influence on the depositional patterns of portions of the middle member Bakken.
6. Significant oil shows scattered throughout the study area suggest a strong potential for successful future exploration.
7. The producing sandstone unit mapped in the study area may have provided the migration pathway for Bakken oils.

ACKNOWLEDGMENTS

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BREAKING INTO BAKKEN POTENTIAL ON THE FORT PECK RESERVATION IN NORTHEASTERN MONTANA

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ABSTRACT

The Bakken Formation is recognized as an important hydrocarbon source rock for the Williston Basin. Recent developments in horizontal drilling technology have made this unique geologic formation a principle exploration target as well. Necessary ingredients for a Bakken play involve overpressuring, thermal maturity, stratigraphic thinning, hydrocarbon content, and especially, the presence of fractures to free the oil. Tectonic breaking of the Three Forks, Bakken, and lower Lodgepole rocks is likely to occur near local and regional tectonic elements associated with Laramide, reactivated, and salt dissolution events. Other Bakken cracking happens during compaction and hydrocarbon expulsion and during erosional pressure unloading.

Bakken thickness varies on the Fort Peck Reservation from 0 to 25 m (0 to 75 ft). The Upper Shale Member is remarkably uniform, ranging from 3 to 4 m (8 to 12 ft) in most places. Stratigraphic thickness appears to be related to a northwest structural grain, especially in the northeast where fold axes are located parallel to the Opheim Syncline. This strike is coincident with the salt edge of the Devonian Prairie Evaporite. The Bakken is about 16 m (50 ft) thick along this dissolution boundary and may be fractured. Structural flexure near the Wolf Creek Nose, and especially off the northeast and eastern flanks of the Poplar Dome, where seismic profiles display steep normal faults, may have suitably fractured the Bakken, as well. Well logs in this area reveal good resistivity separation in the Middle Siltstone Member of the Bakken, and this feature may be used to detect fracturing in the low porosity reservoir rock. In addition, the lower Lodgepole has a similar electrical resistivity response on some well logs. The Three Forks is anomalously thick northeast of Poplar Dome.

Depth below ground of the Bakken varies from about 2100 to 3050 m (6500 to 9500 ft). Electrical resistivities and sonic travel times indicate that much of the Reservation's Bakken lies at depths sufficient to generate hydrocarbons. Resistivity and gamma ray logs confirm that organic carbon content is adequate, if not exceptionally high. Regional Laramide uplift, coupled with glacial erosion and rebound, probably account for the present physical state of the Bakken in this area. Overpressuring probably exists over at least half of the Reservation as determined by sonic log calculations and sparse drill stem test pressures.

INTRODUCTION

The Bakken Formation is receiving considerable attention in northeastern Montana as a potential reservoir for oil. Long known as a prolific source rock for Paleozoic hydrocarbon reservoirs in the Williston Basin, this rich organic shale, when fractured, can produce oil. Successful horizontal drilling and completion of the Bakken in western North Dakota have led explorationists into northeastern Montana. Although, no significant discoveries have yet occurred, leasing activity, horizontal tests, and vertical well production all point to Bakken interest.

This paper summarizes Bakken exploration parameters under the Fort Peck Indian Reservation (see Fig. 1). The study has been prompted by recent, and extensive leasing on the Reservation in at least two Bakken plays. As part of an ongoing Minerals Assessment Project funded by the Bureau of Indian Affairs, this report will provide a partial preview to that evaluation which may be presented in 1993 at a northeast Montana Symposium.

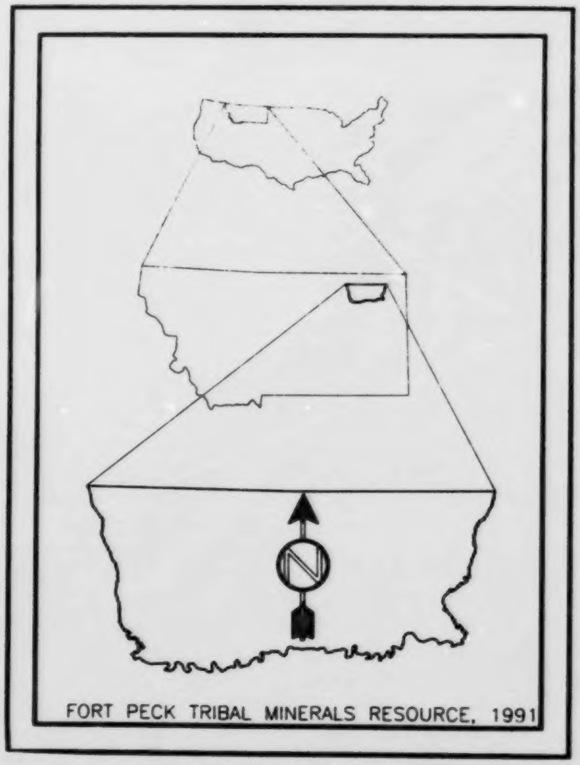

Figure 1: Index map for the Fort Peck Reservation northeastern Montana, U.S.A.

Optimism predominates in the work of several authors who have examined Bakken potential in the Fort Peck Reservation Area. This paper will build on that positive outlook and attempt to confirm previous conclusions without unnecessary duplication of those efforts. The Reservation will be examined on two levels of data coverage. Well information from approximately 1,100 wells was used to construct reservation wide maps using the GeoGraphix Exploration System. This data set includes wells in a one township buffer around the Reservation (T26-34N, R39-56E). Additionally, public domain gravity and magnetic data, and Landsat satellite images were used to construct maps on those respective data sets. 450 mi (726 km) of seismic data was acquired, reprocessed, and interpreted for the eastern third of the Reservation. This dataset was independently contoured for comparison and for better evaluation of areas under leasing activity. The eastern region contains relatively unexplored Bakken and other Paleozoic potential.

HISTORICAL EVIDENCE

The work of several geologists decidedly points to the possibility of Bakken oil production on the Fort Peck Reservation. One of the first and most lasting proponents of the Bakken is Fred Meissner (1978). His, and other Bakken "guide-lines" appear in figure 2. In general, his study (1978) did not extend very far on to the Reservation. He inferred the probability of abnormal high pressure and thermal maturity, in at least the eastern portion of the Reservation. These conditions were based on high sonic travel times and high resistivity in the Bakken shales. Much of his work concentrated on explaining the structural reasons for the conditions which help to create fractured reservoirs. His recognition of the importance of Stearn's (1971) type "drape folds" in Bakken reservoirs, forms a basis for structural observations made in this paper. Previously identified structures are labeled in figure 3.

Schmoker and Hester (1983) extended the Bakken study across the Reservation. Their limits for prospective Bakken thickness, high resistivity, and thermal maturity, lie across the southwestern corner of the Reservation. Implicit in these log-derived boundaries is that most of the Reservation is within potentially productive Bakken. Their conclusions echo this deduction with the words "present formation temperatures are roughly proportional to the level of thermal maturity, resistivity values define the region where oil saturations are high, and density and neutron-log data show that a significant amount of organic matter in the shale members has been converted to oil and expelled".

A statistical model developed by Krystinik and Charpentier (1987) demonstrated that present day organic carbon distribution in the upper Bakken agrees with previous geochemical analysis. As shown in figure 2, their area of high organic carbon in northeastern Montana includes roughly half of the Fort Peck Reservation.

Hansen and Long (1991a) have asked how far the Bakken "Fairway" extends in their recently published contribution to the Montan Geological Society conference proceedings entitled "Geology and Horizontal Drilling of the Bakken Formation". They define the Bakken production trend in North Dakota and its logical extension into Montana by the relationship between the apparent water resistivity (Rwa) and free oil recoveries in Bakken drill stem tests. The resistivity value is calculated in the middle siltstone member, because it has matrix porosity, log characteristics which can be analysed, and is perforated in some vertical wells. In a second paper (1991b) by these authors two exploration criteria that characterize Bakken producers are noted: resistivity separations in the middle siltstone and relatively high, final shut-in pressures in the drill stem tests. Both criteria may indicate the required fracturing necessary to create Bakken reservoirs. They have extrapolated the "fairway" onto the Reservation. Numerous examples of Bakken resistivity separations in Reservation wells have been found.

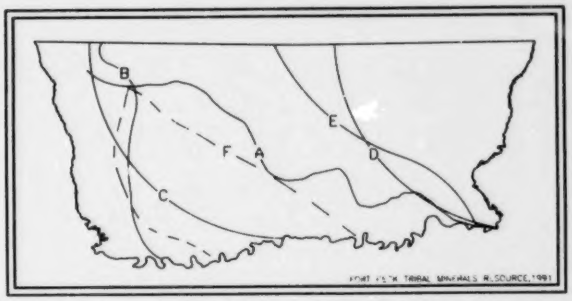

Figure 2: Previously identified Bakken "guide-lines"; A) Limit of high organic carbon content in Bakken shales, Krystinik and Charpentier, 1987); B) Western limit of U. Bakken resistivity > 100 units, Schmoker and Hester, 1983; C) 175°F BHT, western limit of thermal maturity, Schmoker and Hester, 1983; D) Fluid gradient line of .70, western edge of high pressures mapped by Meissner, 1987; E) Prairie salt dissolution edge, Peterson and MacCary, 1987.

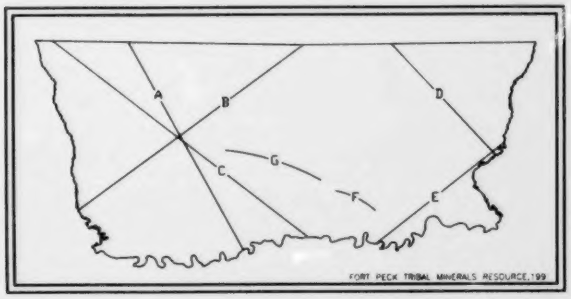

Figure 3: Known geologic structures; A) Cedar Creek Trend, B) Poplar Fault Trend, C) Projection of Bi-centennial Trend, P.L., 1990, D) Opheim Syncline, E) Brockton-Froid Fault Zone, F) Poplar Anticline, G) Wolf Creek Nose. A,B,E from Peterson and MacCary, 1987. D,P,G from several sources.

BAKKEN EXPLORATION PARAMETERS

Breaking into Bakken potential involves careful consideration of the necessary criteria for production in this unique and challenging reservoir.

The following parameters are reviewed:

1) Oil Shows
2) Total Organic Carbon
3) Thermal Maturity
4) Stratigraphic Thinning & Geometry
5) Structural "Cracking"

BAKKEN OIL

A cursory examination of Fort Peck Reservation Area wells reveal that the Bakken does in fact have free oil and gas. Bakken shows are reported in the Reserve and Big Muddy Creek Oil Fields (Ames, 1985; Nagel, 1985). These are two of the many multi-pay Paleozoic oil fields that are adjacent to the Reservation's eastern boundary. Table 1 summarizes drill stem test (DST) data for nine of the 20 area wells in which the Bakken was sampled. The 9 tests generally had the longest shut-in times. Many of the others were shut-in for less than an hour. Several other wells have tests erroneously reported as Bakken instead of the Madison and especially, the Nisku or Duperow formations. Most of the DST's in Table 1 recovered gas; wells 4, 6, and 9 had oil. Hydrostatic pressure in most of the DST's was between 3500 and 5500 psi (24,000 - 38,000 kPa). Four wells tested shut-in pressures greater than 2000 psi (13,790 kPa) and four DST's on Table 1 show an increase in the final shut-in pressure. Fractures may be present in these Bakken reservoirs.

Bakken production from vertical wells occurred in North Dakota long before any horizontal wells were attempted. Two wells which have produced from the Bakken are within 10 mi (16 km) of the Reservation. The Parson 14-27, (T26N-54E-Sec. 27) produced 9,943 barrels of oil and 867 barrels of water in 56 online months between May, 1983, and July, 1989. In August of 1989 the Finnicum 31-44 (T26N-55E- Sec.31) was recompleted in the Bakken and through January, 1991, has produced 7,805 barrels of oil and 617 barrels of water. In 1964 an attempt was made at establishing Bakken production on the Reservation in the E. Paulson 1 well (see table 1). A DST in that well did recover gas cut mud all the way to the surface, however several production tests.were unrewarded.

Table 1: Selected Bakken DST's, Fort Peck Reservation

WELL NAME	TWN	RGE	SEC	DST TOP	DST BASE	ISIP	FSIP	SI	RECOVERY
1.* Soo-Tribal 1	32N	52E	20	8537	8582	3215	3145	.5	160'M
2.* Sethre 1	30N	47E	12	7670	7676	2318	2214	.75	737'SGCZW
3. Beck 1-36	29N	56E	36	10078	10111	151	218	6	980'G 90ft SGCP
4. Munz 1	29N	56E	1	9970	10056	340	2807	3	93' OCMXW
5. Miller/Fed 9-1	26N	54E	9	9290	9364	216	117	4	85' M 14 c.f.g.
6. Land 1-27	32N	56E	27	8790	8880	696	471	2	196'SGCM 475cc O
7.* Tiegan 1	33N	46E	27	7150	7204	1807	2078	4	186' MCW
8.* E. Paulson 1	33N	49E	34	7900	7970	3560	3769	1	360'G 6345 GCM
9. Grimsrud 1-28	34N	56E	28	8534	8600	193	245	5	180'G 50 cc OCM

* Well on Reservation
DST Top and Base are depths, ISIP = initial shut-in (p.s.i.) FSIP = Final...,
SI Time = Final shut-in time (hours).

AN INDIGENOUS SOURCE

Although more and more reservoirs are being identified as having source rocks within the same formation as the reservoir rock, the Bakken is the ultimate source/reservoir combination. With total organic carbon (TOC) analysis greater than 10 percent in many parts of the Williston Basin (Dow, 1974; Williams, 1974), the Bakken is an impressive hydrocarbon factory. The absence of core samples in an area such as Fort Peck necessitates the use of indirect indicators of TOC such as well log gamma ray, resistivity, density, and neutron readings. All were used by Schmoker and Hester (1983) to outline where the Bakken has enough TOC and where the carbon has been liberated by high temperature. Meissner (1978) used resistivity to delineate an area of "mature, high resistivity", in the Bakken Shale. His study did not include much of the Fort Peck Reservation, but his methodology can easily be applied to this area with similar results. The TOC work of Krystinik and Charpentier (1987) has already been cited.

COOKED TO PERFECTION

High TOC does not ensure production. Appropriate thermal maturity must be reached to generate oil. Meissner (1978) set an upper limit of 6,200 ft (1891 m) in depth to generate sufficient hydrocarbons. Plots of several Bakken "guidelines" that have been determined by this study are shown in figure 4. (Note the 6,500 and 8,500 ft present depth contours). Temperature must reach at least 160°F (71°C). Figure 2 has a 175°F (79°C) line that includes most of the Fort Peck Reservation. Erosion and glaciation have removed some of the stratigraphic cover in this area. It is entirely possible that the Bakken has been at greater depths and higher temperatures in the past. Thermal maturity should be more than adequate throughout most of the Reservation.

THE RIGHT STUFF

The stratigraphy of the Bakken plays an important, if not, dominant role in Bakken productivity. It is where the shales are thin that they become more brittle and susceptible to fracturing. The Bakken "Fairway" of western North Dakota lies along the updip depositional (?) edge of particularly the upper shale member. Hansen and Long (1991) have postulated an extension of that stratigraphic trap with a well log technique employing Rwa values less than 0.1 (see figure 2). This extrapolation projects into an area of the Reservation that has very few wells, until the Lustre Oil Field is reached. Each of the Bakken member "pinch outs" are approximated in figure 4. Many Lustre wells have less than 5 ft (1.5 m) of Lower, or even Middle Bakken thickness. The western part of the Reservation lies over the region with less than thermal maturity and thus enough oil may not have been liberated in this area. It should be emphasized that the Bakken is relatively thin across the entire middle region of the Reservation. The 50 ft (15 m) isopach line for the Bakken Formation is plotted on figure 4. Even this far east, the upper and lower shales are usually less than 10 ft (3 m) in thickness and thus thin enough to be easily fractured.

Figures 5 a-d display the Bakken thickness variations and trends. The regional depositional, northwest "grain" is evident in all maps, but each member has individual anomalous behavior. In figure 5a, the Upper Bakken Isopach, thickness patterns vary from a west/northwesterly orientation in the west, to a more northerly direction in the center, and finally to a northeasterly disposition in the east. This and the circular anomalies most likely have different tectonic implications. The Upper Bakken is almost gone over the Poplar Dome. In fact, it is usually less than 10 ft (3 m) in thickness across the Reservation. The middle siltstone member (Fig. 5c) has less variability, and is characterized by the regional northwest orientation. Maximum thickness on the Reservation is 50 ft (15 m). Several contour reversals and the circular anomaly, all in the northeast, may have a structural basis. The Lower Bakken as displayed in figure 5b, mirrors what was observed in figure 5a. The regional northwest depositional pattern is offset by northeast trending thickness patterns. Again, as in the upper shale, the Lower Bakken is usually less than 10 ft (3 m) in thickness with several circular anomalies of double that thickness. The total Bakken Isopach map resembles the Middle Bakken Isopach Map (compare figures 5d and 5c). However, the northeasterly contour pattern observed in both the upper and lower shales appear on the total thickness map.

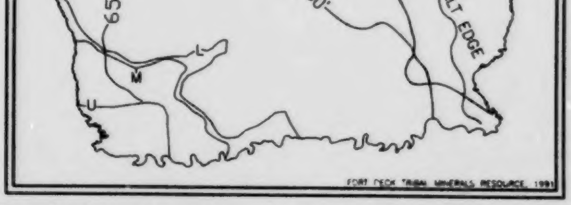

Figure 4: Bakken guidelines, this report. U.M.L - limit of Upper, Middle, and Lower members of the Bakken Formation. 50' - 50 foot total Bakken thickness contour. 6500' - 8500' - present depth contours. Prairie Formation salt edge included.

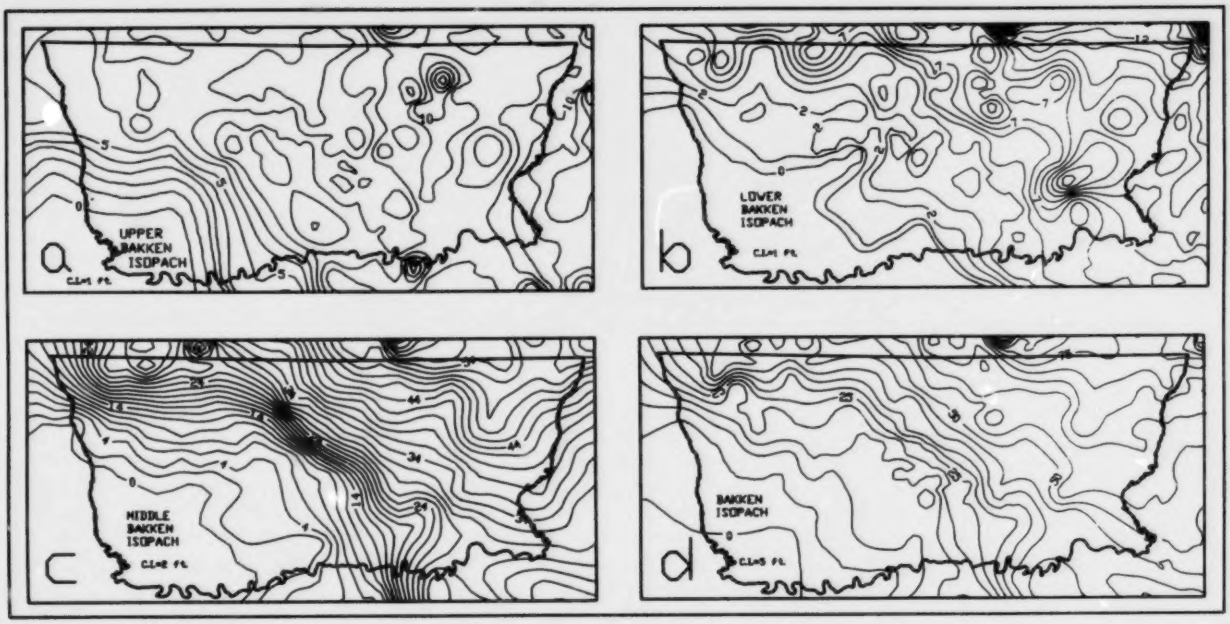

Figure 5: Bakken Formation isopach maps for: a) Upper Shale; b)Lower Shale; c) Middle Siltstone; and d) Total thickness.

AT THE BREAKING POINT

Breaking into Bakken potential anywhere requires just that; fracturing this organic rich, thermally mature, and depositionally thin, shale. In western North Dakota, regional tectonic fracturing and thermal, oil-expulsion cracking, are apparently enough to create a fractured Bakken reservoir. "Sweet Spots" are found probably where local tectonic impetus has been added. The Bakken also produces along major tectonic elements such as the Nesson Anticline. Tectonism serves a dual purpose by controlling Bakken thinning over pre-Bakken structural highs and by stressing the lithified Bakken shales, particularly by Laramide and even post-Laramide movement in this area.

Figure 3 features some regional structural elements that others have long recognized in the Fort Peck Reservation area. The Basin-wide, northwest grain, with subordinate northeasterly elements is obvious. Figure 6 is a summary of key structural fold axes determined from contour maps of well data and seismic information. The Tri-county, Smoke Creek (north and south), and North Lobo folds are new names proposed in this paper. The Opheim Syncline has not been traced in such detail before, either. The present day Bakken subsea structure is presented in Figure 7. Seismic time contours give detail to the sparsely drilled eastern third of the Reservation (see map inset).

The faults flanking the Poplar Anticline (Fig. 6) were detected in seismic profiles and are vertical offsets of small displacement that relieved stress on the Poplar building event. They often die out in stratigraphic units near the top, or immediately above, the Paleozoic. They are significant because they confirm the tectonic drape fold model proposed by Stearns (1971) and alluded to by Meissner (1978) for the Antelope Field in North Dakota. This model forms the basis for a lineament block analysis that Shurr, Monson, and Lund have recently presented and expanded (1991). (see figure 8 for a summary of that analysis). The brittle Paleozoic carbonates, and Bakken shale for that matter, are faulted by recurrent crustal block movements that both control deposition and fold the overlying plastic Mesozoic clastic units. Reversal in movement has been documented in at least one fault that does propagate through to the Eagle horizon. There is excellent correlation in orientation and proximity between surface lineaments observed on Landsat imagery, and the faults which have been identified. Shurr *et al.*, (1991) believe that the Lineament Blocks outlined in figure 8 represent tectonic zones which contain similar crustal elements that have both vertical, as well as strike slip, senses of movement. Regional depositional patterns in the Bakken relate to the northwesterly blocks found in the central Reservation area, whereas anomalous, northeasterly trends in the upper and lower Bakken shales have been influenced by northeasterly trending tectonic elements. Block rotation at lineament zone intersections has also influenced depositional thickness and may have deeper crustal causes. The Lineament Zone intersection in the northeast in figure 8 is particularly interesting, in that numerous stratigraphic units including the Bakken siltstone, have circular thickness anomalies at this location. A connection to crustal elements is strongly suggested by a magnetic anomaly here called the Smoke Creek Anomaly (Shurr, Monson, and Lund, 1991). Seismic time contours, and the Prairie salt edge (Fig. 4) are also deflected in this location.

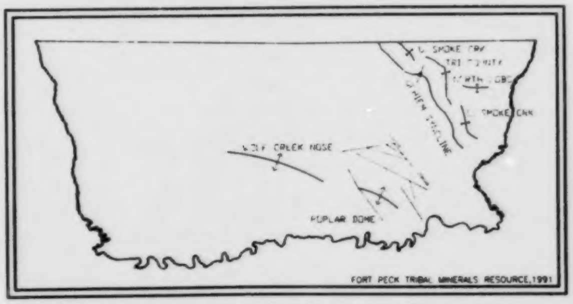

Figure 6: Reservation structures, this report. Faults flanking Poplar Dome are predominently high angle normal faults with upthrown block towards Poplar structure.

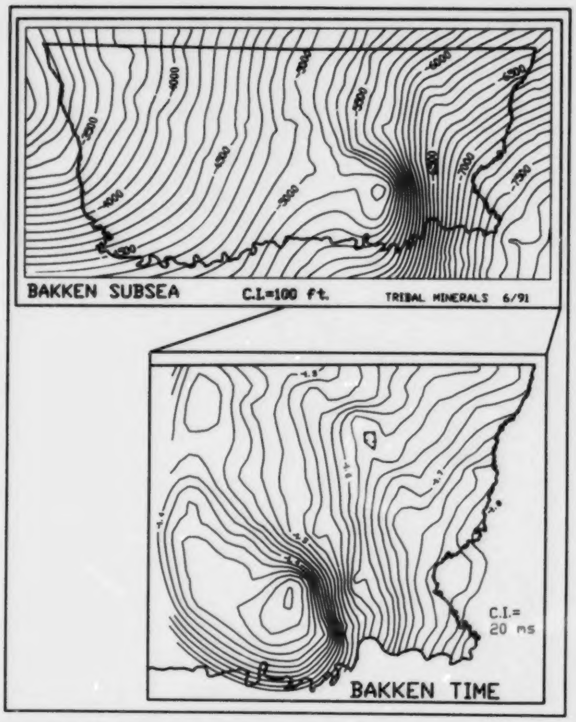

Figure 7: Bakken subsea and seismic time structure maps. Time map is for eastern third of Reservation.

A more thorough explanation of the tectonic history will be prepared as part of the Reservation's ongoing Minerals Assessment Project. That paper will compare lineament analysis, gravity, magnetic, seismic, and well data. For now it is sufficient to outline what appears to be the tectonic framework beneath the Fort Peck Reservation.

The picture is clearest in the central to eastern part of the Reservation. Geophysical data suggests that three tectonic regimes exist:

1) The central Reservation block
2) A transitional zone
3) An Eastern, Williston Basin Element

These three regimes are identified in figure 9. The west or Central Regime is an elevated, low relief, structural element that has been relatively high at different times affecting certain Paleozoic carbonate and Cretaceous sandstone deposition. Most of the Reservation's oil production was found here. The Transition Zone marks a steeper structural gradient which "rolls" over into the Opheim Syncline. An even more noticeable dip reversal occurs between the Transition Regime and the Central Reservation Block (see Fig. 9b). Faulting is more likely in this regime. The Opheim Syncline axis is a critical line in the eastern part of the Reservation. Structural maps in figure 10, are contoured on two horizons: the Cretaceous Eagle Formation and the Ordovician Winnipeg Formation. Seismic time maps for each horizon are compared in figure 11 and add important delineation of structure. The parallel anticlines, associated with the Opheim Syncline in the northeast, are sympathetic flexures that also may have fractured the Bakken. The Poplar Lineament Zone probably is bounded by the Opheim tectonic regime line. There is, however, no apparent seismic offset across this boundary, nor across the famous Brockton-Froid Fault.

When working with seismic data it is often helpful to use isochron maps as indicators of tectonic elements, and thus the stress environment that might fracture the Bakken. Two such isochrons are illustrated in figures 12 and 13 for the Piper to Bakken, and Bakken to Winnipeg intervals, respectively. Cross section insets for each isochron map display structure at Piper and Bakken time, respectively. The contrast is dramatic, especially in reference to the Opheim Syncline. The pre-Bakken interval appears to follow the fold axis exactly suggesting that deposition was related to this structural hinge line. For the post-Bakken interval, isochron thickness changes at an angle perpendicular to the fold axis. Isochron patterns for each interval are similar in the west on the central Reservation block; *e.g.*, both thin over the Poplar Dome, which means the Poplar structure was active even in the Paleozoic. The more rapid change in post-Bakken thickness on the eastern flank of the Poplar dome points to more rapid growth and perhaps more extensive fracturing of the Bakken during this time. The Prairie Evaporite edge is prominently displayed in figure 4 and again in figures 12 and 13. Thought by

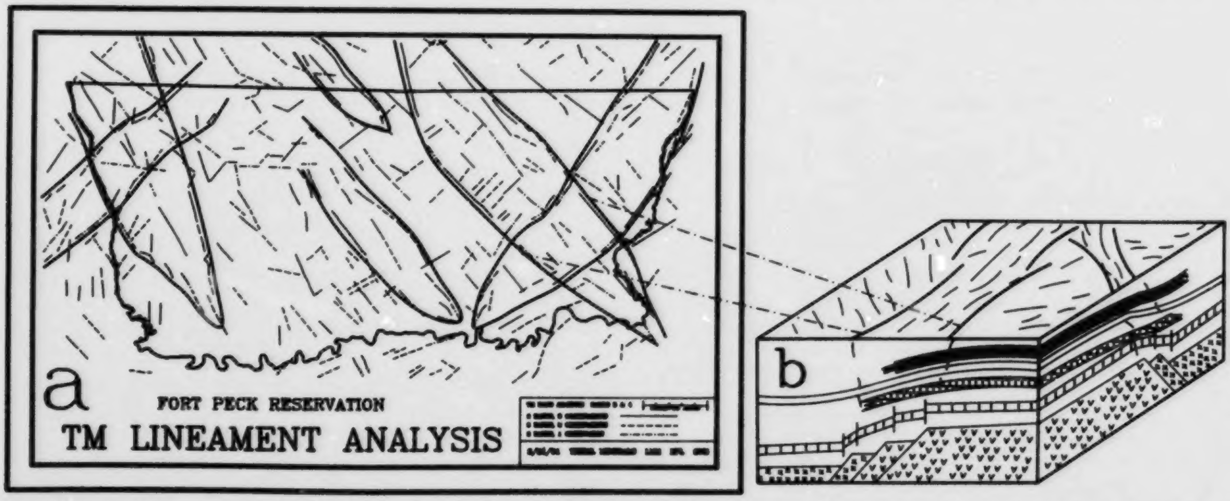

Figure 8: Satellite Lineament Analysis; a) Observed lineaments from TM bands 3 and 4. Dark heavy lines outline zones of common tectonic elements; b)Block model diagram. (Shurr, Monson, and Lund, 1991).

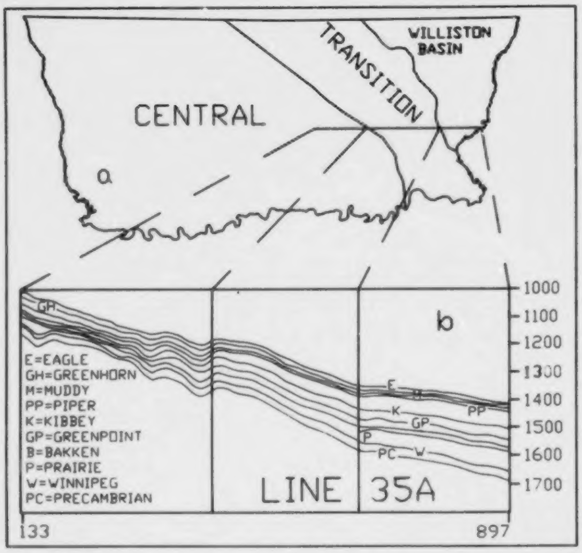

Figure 9: Tectonic regimes on the Fort Peck Reservation.; a) See text for explanation.; b) Seismic event profile for line 35A.

many to be a salt dissolution limit, this linear feature has important ramifications in defining Bakken exploration parameters. Swenson (1967) placed the original depositional limit of the Prairie across the center of the Reservation. He was convinced that thickening in the Dawson Bay carbonate followed early stage dissolution channels, and that later, widespread salt removal and subsequent compaction, created small dolomitized structures in the Nisku (Birdbear) Formation. Rogers and Mattox (1985) neatly outline the two seismic isochron parameters necessary to confirm Prairie dissolution: 1) Bakken to Winnipeg thinning under; and 2) Piper to Bakken thickening. Such salt solution tectonics would most likely stress the Bakken shales and fracture them. This is most likely the guiding model for recent leasing and horizontal well tests by at least two operators in northeastern Montana.

Evaluation of seismic data has permitted the following observations:

1) As seen in closely spaced wells where the Prairie is 100 ft thick in one well and absent in the other, seismic profiles show an abrupt termination of the salt.

2) The edge is considerably more sinuous than mapped by Sandberg (1961) or Peterson and MacCary (1987).

3) The salt boundary closely mimics the Opheim Syncline axis, usually lying to the east (basinward) of it. Good congruency of strike exists with respect to the Lower Paleozoic strata (see Fig. 13, Bakken to Winnipeg isochron map).

4) The Prairie salt surface is parallel to the Bakken Formation as seen in the profile in figure 13. If projected west it would intersect an already elevated Poplar Dome.

5) There is no compensational thickening in the rocks in the region beyond the Prairie Evaporite.

These observations, especially the last two, suggest that the Prairie edge may be depositional. The three part tectonic regime model proposed earlier supports this if the area to the west has always been relatively higher. Erosion shortly after deposition is also a possibility. In any event, folding is observed on either side of the Prairie edge. Whether or not these dip reversals are due to salt solution or tectonic flexing along the Opheim Syncline, is irrelevant as to the state of fracturing in the Bakken. Present day structure can be accounted for by either agent acting on the Bakken.

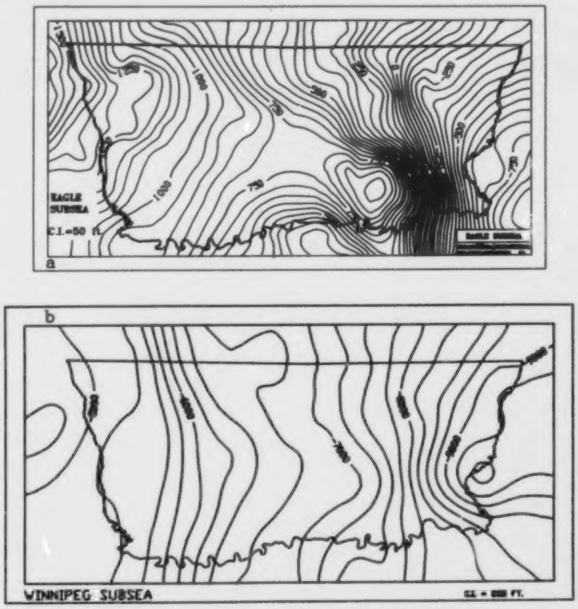

Figure 10: Structural contour maps on the Cretaceous Eagle and Ordovician Winnipeg Formations.

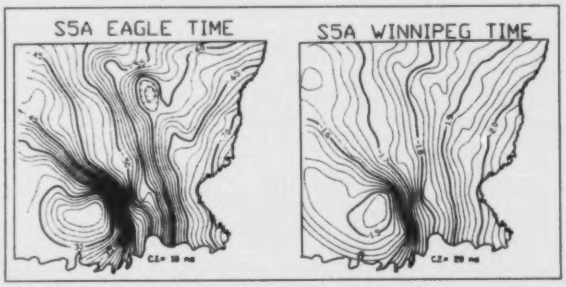

Figure 11: Seismic time contour maps for the eastern third of the Reservation. Eagle and Winnipeg markers.

CONCLUSIONS

On the Fort Peck Reservation the Bakken Formation provides an excellent hydrocarbon exploration opportunity. The upper and lower shales are sufficiently thin to be easily fractured. All evidence suggests that the high levels of organic carbon have been adequately buried and heated to generate hydrocarbons. Although significant overpressuring to expel the resource into fractures has not been verified by the few, short DST's that have been run, a projection of Meissner's abnormal pressure gradient zones into the region is likely. Ample local tectonic fracture possibilities exist in and around the Wolf Creek Nose and Poplar Anticline, the faulting northeast and east of Poplar, and along either side of the Opheim Syncline. Here the Prairie salt edge, the structural reversals associated with it, and the parallel anticlines to the east, offer excellent possibilities. The regional tectonic fabric, as

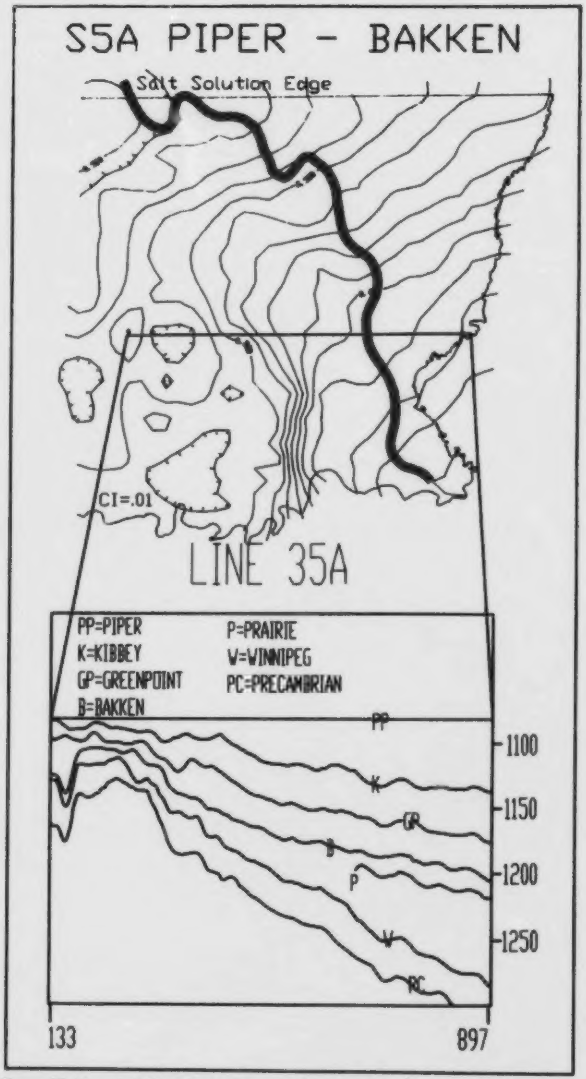

Figure 12: Piper to Bakken isochron map with a seismic horizon profile flattened on the Piper event. Note Prairie salt edge.

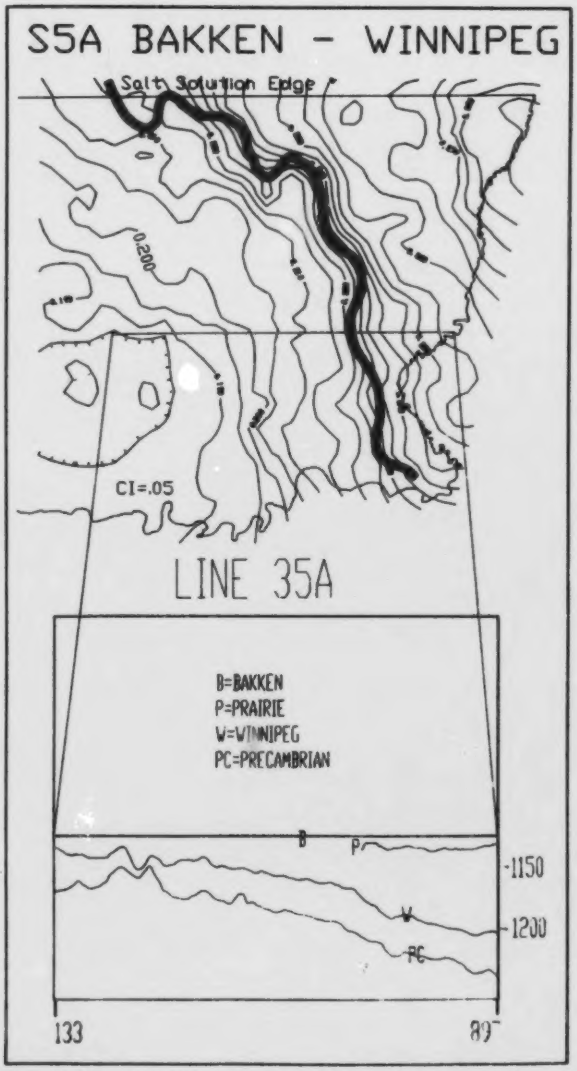

Figure 13: Bakken to Winnipeg isochron map with a seismic horizon profile flattened on the Bakken event. Note Prairie salt edge.

delineated by satellite lineament zones, may offer some criteria for fracture intensity determination and orientation. Finally, the stratigraphic and depositional patterns acting in response to these structural features can be guides to Bakken exploration targets.

ACKNOWLEDGEMENTS

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ROUTLEDGE POOL: A MODEL FOR DEVELOPMENT OF THE VIRDEN-WHITEWATER LAKE PALEOEROSIONAL TREND, SOUTHWEST MANITOBA

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ABSTRACT

The Virden-Whitewater Lake paleoridge on the Mississippian erosion surface is located on the northeastern margin of the Williston Basin in southwestern Manitoba. In the Routledge pool area (Township 9; Range 25 W1) along the northern portion of this paleoerosional feature, Mississippian Lodgepole strata subcrop across a narrow northwesterly trending paleoerosional high, underlain by parallel basement-controlled structural elements. Oil is trapped in Lodgepole oolitic and skeletal carbonate sand reservoirs beneath dense dolomite caprock and against a steep updip escarpment facing northeast. Beyond this cuesta deep erosion into Upper Devonian rocks is seen within a short distance.

The paleoerosional high across the Routledge pool area can be traced as a continuous element for at least 110 km (70 mi), creating a paleodrainage watershed to the southwest into the main Williston Basin depocentre and to the northeast into the deep Brandon sub-basin. Erosion backstepped progressively southwestward towards the escarpment without secondary erosional forms and left no apparent outliers. The Virden-Whitewater Lake ridge bears an obvious relationship to basement-controlled structural elements. A major stratigraphic accumulation of oil in the Virden and Routledge pools straddles the ridge and is sealed up-dip against tight dolomitized Lodgepole strata.

Beyond, to the south the Virden-Whitewater Lake trend is traced discontinuously for at least 55 km (35 mi) to the international border across outliers of Lodgepole strata. Breaching of the ridge does not coincide with apparent underlying structural control. Oil is trapped in Whitewater Lake Member reservoirs preserved within outliers overlain by Lower Watrous siltstone and shale.

INTRODUCTION

The overall objective of this study is to provide a detailed description of a unique Mississippian unconformity capped oil pool, and to present a model for genesis of the Virden-Whitewater Lake trend.

In the Routledge pool area of southwestern Manitoba, oil is trapped in subcropping Mississippian Lodgepole limestone reservoirs against paleo-cuestas facing northeast counter to regional Williston Basin dip (Fig. 1). Behind the front escarpment is a northwest-southeast trending ridge on the Mississippian erosional surface, across which progressively older Lodgepole strata subcrop in an up-dip direction. Paleoerosional relationships seen at the Routledge pool can be traced for over 150 km along the Virden-Whitewater Lake trend.

The pool study area is confined to Townships 7 to 10; Ranges 24 and 25 W1, (Figs. 1, 2) and includes the Routledge and East Routledge pools (Township 9; Range 25 W1) and the eastern portion of the Virden-Roselea pool (Township 10; Range 25 W1). Discussion of a regional model for development of the Virden-Whitewater Lake trend from the Manitoba-Saskatchewan border at roughly Township 15, to the Manitoba-North Dakota border at Township 1; Range 20 W1, follows.

PREVIOUS STUDIES

After discovery of the Virden pool (Fig. 1) in 1953, a number of preliminary geological studies were published by Atkinson *et al.* (1956), Berg (1956), Milne *et al.* (1955), Owen (1953) and Stanton (1955, 1958). Regional studies of the stratigraphy and oil pools along the Virden-Whitewater Lake trend by Manitoba Energy and Mines (McCabe, 1959; 1963; Ferr *et al.* 1987; Martiniuk *et al.* 1986; Martiniuk, 1988; 1989) have contributed significantly to this paper. In addition, Young (1973) and Young *et al.* (1975) described regional Lodgepole stratigraphy and diagenesis with relationships to oil accumulations.

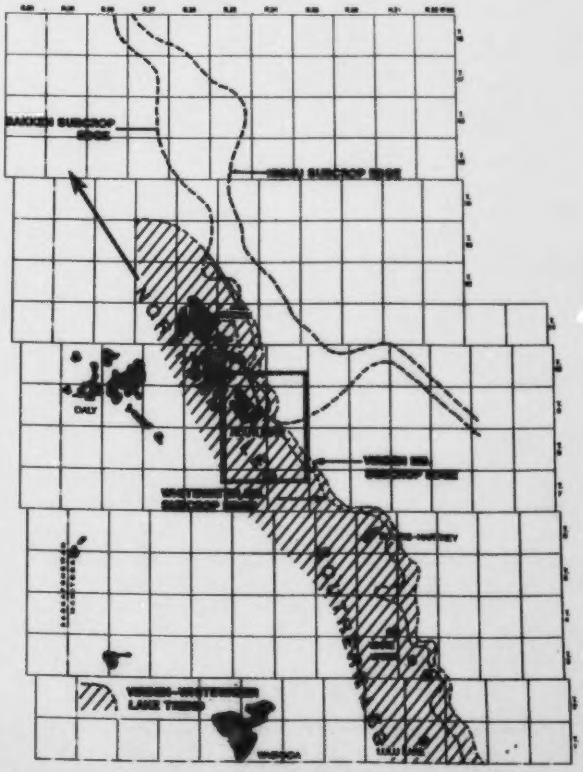

Figure 1: Index map showing location of Virden-Whitewater Lake trend relative to major subcrop edges, productive areas, and Routledge pool map area.

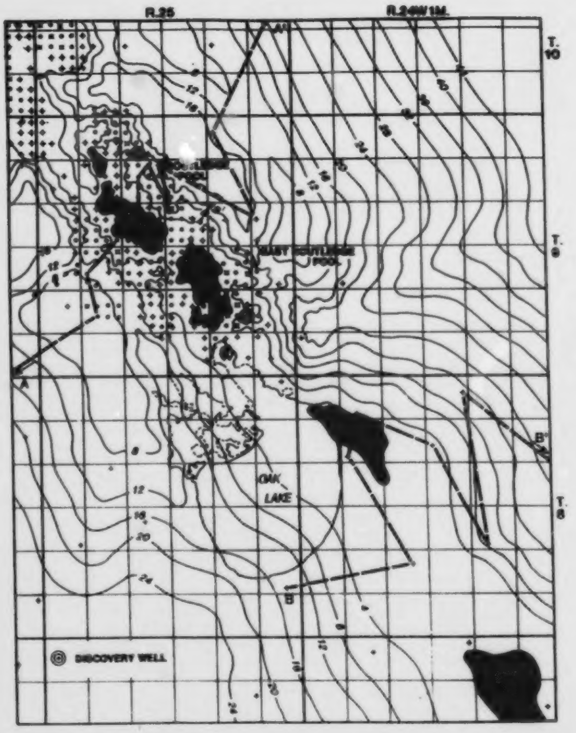

Figure 2: Isopach map, Lower Watrous Red Beds, Routledge Pool area; 2 m (6.5 ft) contour interval; stippled area less than 2 m isopach. Cross sections A-A' and B-B' indicated.

PRODUCTION HISTORY OF THE ROUTLEDGE POOL

The Routledge pool was discovered in May, 1955, at Lsd. 3-33-9-25 W1 (Fig. 2), near the up-dip edge of the pool in the Lodgepole Formation at roughly 630 m (2080 ft) depth. Pool development on 40 acre spacing units proceeded to 1964 with a total of 146 oilwells. These produced 2,034 x 10^3m^3 (12.81 million bbls) of oil to November, 1990. Presently, there are 83 oilwells in the Routledge pool producing at an average rate of less than 1.42 m^3/day (9 bbl/day). The East Routledge pool discovery (Fig. 2) occurred in November, 1963, at Lsd. 7-11-9-25 W1 and at a similar depth. The pool was fully developed by 1967 on a 40 acre spacing unit with a total of 33 oilwells. In 1972 a pressure maintenance program was initiated using existing oilwells. To November, 1990, total production from the East Routledge pool was 1471 x 10^3m^3 (2.97 million bbls); current average production rates are 1 m^3/day (6 bbl/day). Oil gravity in both pools ranges from 33 to 36° API with very low GOR ratios.

STRATIGRAPHY OF THE LODGEPOLE FORMATION

Subcropping beds of the Lodgepole Formation in the Virden-Whitewater Lake Trend can be subdivided into the following three well-defined stratigraphic units from top downwards (Fig. 3):

WHITEWATER LAKE/FLOSSIE LAKE MEMBER

Subcropping across most of the erosional surface of the Virden-Whitewater Lake trend in the Routledge pool area are the Flossie Lake and underlying Whitewater Lake Members. The Flossie Lake Member comprises dense recrystallized dolomite and limestones with abundant white anhydrite. The Whitewater Lake Member consists of an upper unit of mixed oolite and skeletal wackestones to packstones interbedded with lime mudstones up to 11 m (34 ft) thick and a lower unit of red shale and argillaceous lime mudstone known as the Roselea Shale (McCabe, 1959). Along the downdip edge of the Routledge pool the Whitewater Lake Member has minor oil production potential.

VIRDEN MEMBER

Immediately beneath the Roselea Shale is a 4 to 7 m (13 to 23 ft) thick skeletal and crinoidal sand bed (packstone to grainstone) known as the Upper Virden Member (McCabe, 1959). This lime sand bed, with good intergranular and intercrystalline porosity, forms a consistent major oil reservoir along the northern leg of the Virden-Whitewater Lake Trend and has potential reserves of 12,500 to 79,000 m^3 (80,000 to 500,000 bbls) oil per 40 acre spacing unit. Where the overlying Lodgepole strata are eroded to within 15 m (50 ft) of the Upper Virden Member, dolomitization and anhydrite-infilling obscure primary fabrics and intergranular porosity.

Beneath the Upper Virden reservoir bed is a 12 to 15 m (40 to 50 ft) thick Lower Virden sequence of fine grained white oolitic and skeletal sand beds (packstone-grainstone) interbedded with red argillaceous lime mudstone and shale. Sharp boundaries between each oolite and shaly interbed, together with consistent bed thicknesses, suggest depositional cyclicity upon a shallow marine shelf. In total, there are up to five oolitic sand beds in the Lower Virden, each up to 1.5 m (5 ft) in thickness and of roughly equivalent thickness to the shale interbeds. An exception to this pattern is seen in the East Routledge pool and portions of the Routledge pool, where 6 to 8 m (20 to 26 ft) of thick skeletal sand bed known as the Sandhill lentil (McCabe, 1959) occur (Fig. 3). Dolomitization beneath the Mississippian erosion surface has eliminated the reservoir potential of the Sandhill lentil throughout the Routledge pool area and the adjacent Oak Lake area.

In either direction along the Virden-Whitewater Lake trend from the Routledge pool area, the Lower Virden becomes primarily argillaceous lime mudstone and shale with fewer thin, dense skeletal and oolitic sand beds. Nowhere within this more argillaceous facies do these lime sand beds form an oil reservoir. In the Routledge pool, Lower Virden reservoirs with fair intergranular porosity have reserve potential of up to 15,500 m^3 (100,000 bbls) oil per 40-acre spacing unit. There is no Lower Virden oil production in the East Routledge pool due to porosity reduction caused by dolomitization beneath the Mississippian erosion surface.

SCALLION MEMBER

In contrast to the overlying cyclical shallow shelf depositional sequence of the Lower Virden, the Scallion Member (Fig. 3) comprises white to pink lime mudstone with lesser skeletal or crinoidal wackestone to packstone. White chert nodules, chalky limestone, dense cherty lime mudstone and silicified skeletal wackestone and packstone are common. From the Souris Hartney pool (Twp 6; Rge 22 W1) to beyond the Virden pool (Fig. 1) an 8 to 10 m (26 to 33 ft) thick highly recrystallized and solutioned crinoidal wackestone to packstone lies immediately beneath the lowermost oolite bed of the Lower Virden Member (Fig. 3). This Upper Scallion reservoir bed is best developed within the general limits of the oolitic sand facies of the Lower Virden Member.

When unaffected by dolomitization and anhydrite-infilling associated with the erosional unconformity, the Upper Scallion has good to excellent intercrystalline/intergranular porosity plus solution-enhanced vuggy porosity. Reserves of from 15,500 to 150,000 m³ (100,000 to 1,000,000 bbls) oil per 40-acre spacing unit are present.

In the Routledge pool area a basal argillaceous limestone bed up to 30 m (100 ft) in thickness known as the Routledge shale (McCabe, 1959) lies directly upon the Upper Bakken shale. There appears to be no relationship of this shaly facies to the overlying Lodgepole carbonates. However, from the cherty lime mudstone of the Scallion Member, upwards through oolitic and skeletal sands of the Virden Member, a shallowing-upwards prograding sequence is present along the Virden-Whitewater Lake trend.

UNCONFORMITY SURFACE, ROUTLEDGE POOL AREA

The paleotopographic Virden-Whitewater Lake trend crosses the Routledge and East Routledge pools northwesterly as a ridge up to 5 km (3 mi) in width and 15 m (50 ft) in height (Fig. 2). Lying unconformably upon subcropping Lodgepole strata are Lower Watrous red siltstones and shales (Figs. 2, 3). Less than 2 m (6 ft) of red beds cover the top of the ridge across the central portion of the Routledge and East Routledge pools (Fig. 4, cross-section A-A'), but thicken down the flanks of the high to roughly 10 to 12 m (33 to 40 ft) along the Routledge and Virden-Roselea pool boundaries (Fig. 2).

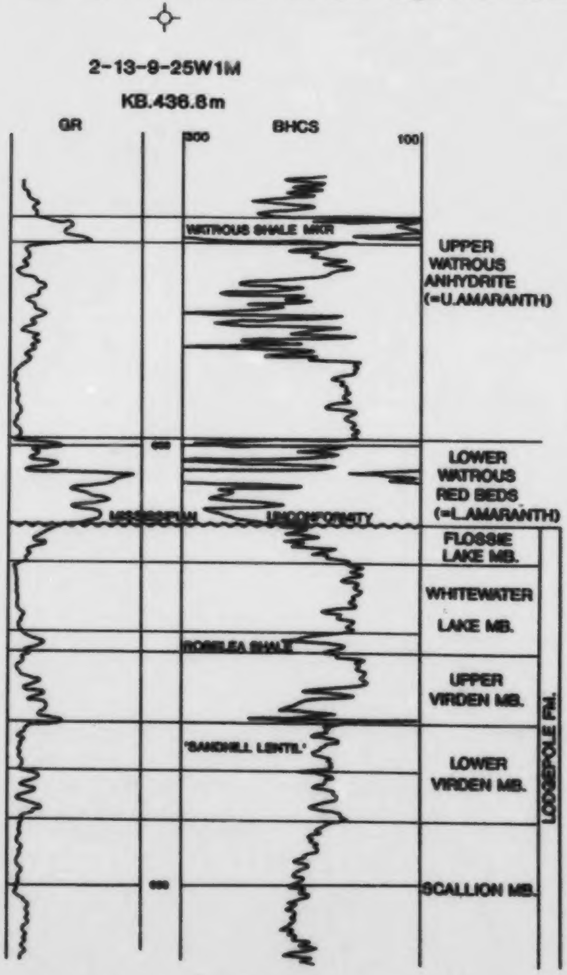

Figure 3: Generalized stratigraphic column, Routledge pool area.

In the Oak Lake area to the southeast, the Lower Watrous likewise increases from less than 2 to 3 m (7 to 10 ft) on the ridge, which here is roughly 4 km (2.5 mi) in width (Fig. 2; Fig. 5, cross-section B-B'). Closure is seen as a broad flat with thin Red Beds centred on Section 29-8-24 W1. As in the Routledge and Virden-Roselea pools to the northwest, the Red Beds thicken dramatically at the up-dip edge of the ridge and beyond into a deep topographic basin to more than 40 m (130 ft) (Fig. 2; Fig. 5, cross-section B-B'). Nowhere in the region does this basin open across the ridge into the Williston Basin depocentre.

Coincident with the Lower Watrous Red Beds thins over Mississippian erosional highs in the Routledge pool study area is thinning of the Upper Watrous Anhydrite (Figs. 4, 5 and 6). As with the Red Beds, this unit thickens to 8 m (26 ft) along the Routledge and East Routledge pool boundaries (Fig. 6). In part this may be due to simple compaction of anhydrite beds. According to Warren (1989), dewatering of a thick gypsum bed causes a decrease in sediment volume and thickness as it compacts with a 38 percent loss of water. Also, compaction of the siltstones and shales of the Red Beds within the lows of the erosion surface may account for some thickening of the overlying beds, as would contemporaneous faulting of underlying Lodgepole strata.

The Upper Watrous Anhydrite shale marker dips southwesterly towards the Williston Basin depocentre at a rate of roughly 3.5 m/km (11.5 ft/mile) in the Routledge pool area (Fig. 7). Structural closure is seen along the northeastern edge of both the Routledge and East Routledge pools. Coincident with the Watrous Beds thick up-dip of the Routledge pool (Figs. 2, 6) is a slight structural reversal on the shale marker (Fig. 7). Counter to the northwesterly structural trends of the unconformity surface, structure contours on the upper shale marker indicate northeasterly anticlinal noses in both the Routledge and East Routledge pools. A significant low at Lsd. 10-15-9-25 W1 between the two pools (Fig. 7) lies astride the paleoridge, thereby (Fig. 2) confirming late fault movement. In fact, this structural low can be traced up-section to above the Cretaceous Second White Specks at approximately 275 m (900 ft) depth. There the displacement is up to 20 m (65 ft) versus the 23 m (75 ft) on the Scallion Member.

To the southeast in the Oak Lake area, the more than 40 m (130 ft) thickening of the Lower Watrous Red Beds at Lsd. 1-13-9-24 W1 is matched by a 28 m (92 ft) thickening of the Upper Watrous Anhydrite (Fig. 6). Structure on the Upper Watrous Anhydrite shale marker rises steadily to the northeast across the deep paleoerosional low (Fig. 5, cross-section B-B'; Fig. 7), without a direct relationship to the latter.

SUB-MISSISSIPPIAN UNCONFORMITY: STRUCTURAL RELATIONSHIPS

To better understand the relationship of structure to the paleotopography, structural contour maps on the Upper Bakken Shale and Scallion Member (Figs. 8, 9) are compared with the maps previously discussed.

Beneath the production zone of the Scallion Member in the Routledge pool area, and beyond into the Oak Lake area, the few deeper well penetrations are insufficient for forming a clear

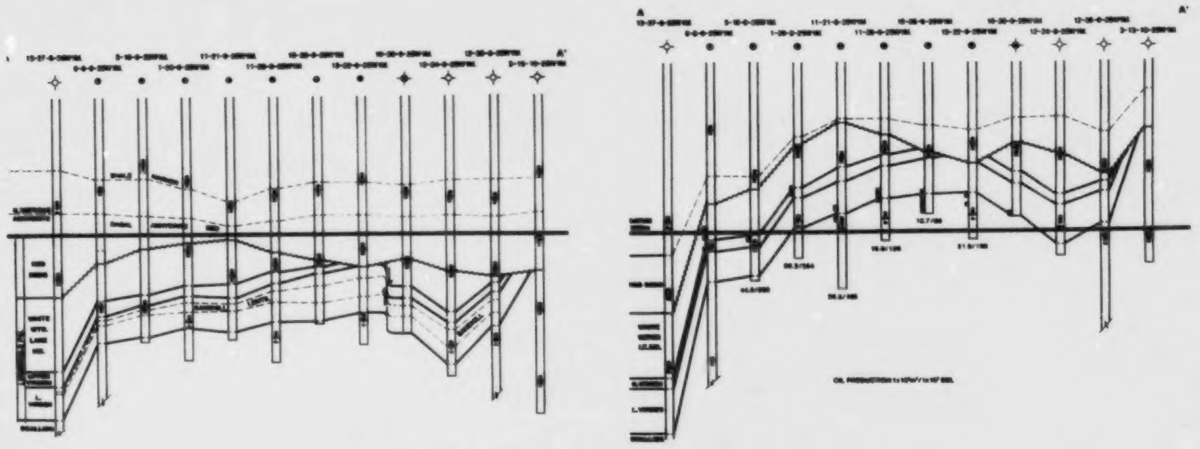

Figure 4: Cross section A-A', Routledge pool area.
a) Stratigraphic section (Lower Watrous Red Beds datum)

b) Structural section (-200m subsea datum). Refer to Fig. 2 for line of section.

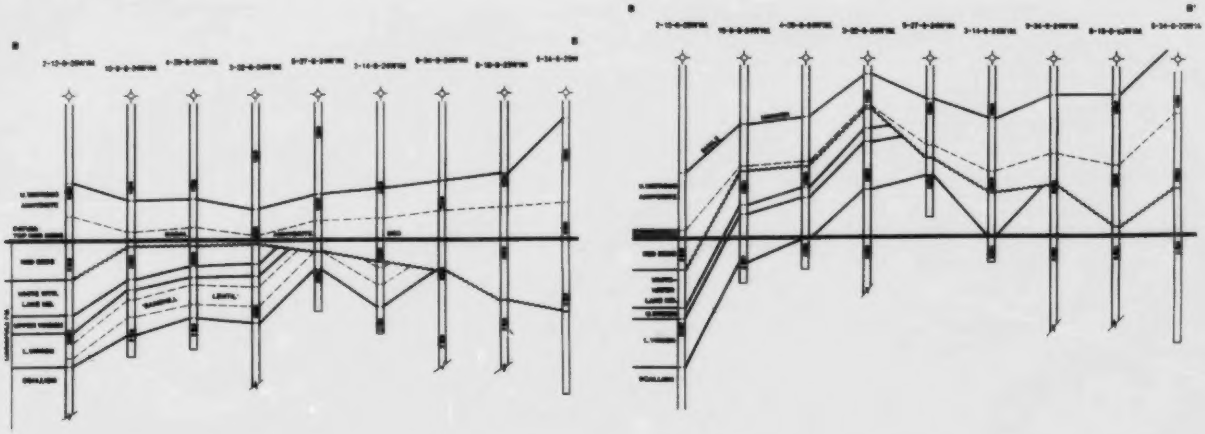

Figure 5: Cross-section B-B', Oak Lake area.
a) Stratigraphic section (Lower Watrous Red Beds datum)

b) Structural section (-200 m subsea datum). Refer to Fig. 2 for line of section.

underlying structural picture. A structure map of the Upper Bakken Shale (Fig. 8a) indicates a strike parallel to the axis of the Virden-Whitewater Lake trend. Along the down-dip edge of the topographic feature, structural gradient on the Upper Bakken increases significantly up-dip (Fig. 8a), and then flattens somewhat over it and beyond to the up-dip edge. This Bakken structural form is illustrated in Figure 8b, a cross-sectional profile from southwest to northeast across the Oak Lake area of the strata between the Upper Watrous Anhydrite shale marker and the Bakken, and where eroded, the Nisku Formation. By assuming a flat Upper Watrous datum, the underlying Bakken and Nisku are shown to rise steadily up-dip beyond the Virden-Whitewater Lake trend, without significant influence on the Mississippian unconformity.

A number of Bakken structural closures on a north-northeast trending nose is indicated by Martiniuk (1988). These include the Waskada oil pool (Township 1, Range 25, 26 W1) and the Hartney structural feature (Township 5; Range 24 W1) (Anderson, 1980). Another would be the 30 m (100 ft) Bakken closure at Lsd. 7-25-7-24 W1 in the Oak Lake area and coincident with the paleotopographic crest (Fig. 8a). It is noteworthy that all closures on the Bakken structural nose roughly follow the north aligned Precambrian Churchill-Superior structural boundary (Gerhard *et al.*, 1982).

Structure on the Scallion Member (Fig. 9) within the Routledge pool area resembles in part that seen on the Upper Watrous shale marker (Fig. 7). In the East Routledge pool, there is a northeasterly trending nose up to 1.5 km (1 mi) in width and up to 15 m (50 ft) high (Fig. 9). It is flanked up-dip by a steep decline of up to 25 to 30 m (82 to 100 ft) between Lsd. 16-14 and Lsd. 4-24-9-25 W1, a distance of less than 600 m (650 yards). In the larger Routledge pool, a similar depression fronts the northwest-southeast antiform between Sections 5-10-25 W1 and 22-9-25 W1 (Figs. 4, 9). As in the East Routledge pool, dip aligned northeasterly orthogonals cross the pools from boundary to boundary (Fig. 9), thereby suggesting two perpendicular structural elements. The structural low separating the Routledge pools at Lsd. 10-15-9-25 W1 is post-Watrous in age (Figs. 7, 9).

A northwesterly trending structural high on the Scallion Member extends from the Routledge pool through the Oak Lake area (Fig. 9). It is truncated by a post-Scallion valley at Lsd. 13-6-

9-24 W1 to within 0.5 km (0.3 mi) of the axis of the structural high. Deep erosion into the Scallion member at Lsd. 9-34-8-24 W1, (Fig. 5, Cross-section B-B'), represents a loss of up to 40 m (130 ft) of stratigraphic section. The Bakken structural closure seen at Lsd. 7-25-7-24 W1 is also reflected by Scallion closure of up to 20 m (65 ft), and is similar to that in the East Routledge pool at Lsd. 8-10-9-25 W1.

An isopach map of the strata from the top of the Mississippian to the Upper Virden Member (Fig. 10) demonstrates dramatic thinning of the Scallion in the Routledge and East Routledge pools. In fact, the region roughly coincident with the 6 m isopach contour of the Lower Watrous Red Beds (Fig. 2) is that where the Upper Virden crops out at the unconformity (Figs. 4, 10). Thinning is also seen along the down-dip edge of both Routledge pools centred upon Section 9 and 10-9-25 W1; in fact, at Lsd. 8-10-9-25 W1, the Upper Virden rises to within 2 m of the unconformity surface. Isopach thicks are seen immediately up-dip of the Routledge pools, and are coincident with structural lows mapped on the Mississippian erosion surface and the Scallion Member (Figs. 9, 10).

In the Oak Lake area, the Upper Virden Member rises to the unconformity at Lsd. 3-14-9-25 W1 within the area delineated by the 6 m Lower Watrous Red Beds isopach as in the Routledge pools (Figs. 5 and 10). From the limited well control it appears that the relationship of the Mississippian erosion surface to the Upper Virden across the Virden-Whitewater Lake trend in the Oak Lake area, resembles that seen in the Routledge pool area.

Another means of relating structural control of the Lodgepole Formation to this complex unconformity is seen in the isopach map of the beds from the top of the Lower Watrous Red Beds to the Upper Virden Member (Fig. 11). Such an isopach interval, straddling an erosional unconformity, can be used to represent structure on the Upper Virden at the end of Red Beds time, by assuming a flat regional datum. An element of error in an interpretation is the assumption of compaction in the thickness of the Lower Watrous siltstones and shales. As a result, this isopach interval, or pseudo-structure map (Fig. 11) throughout the Routledge pool area, demonstrates structural symmetry on either side of the Virden-Whitewater ridge. Scallion structural highs (Fig. 9) in the Routledge pools are similarly reflected on this map, but the highest topographic elevation shown by thinning of the Lower Watrous Red Beds (Fig. 2) is not coincident with original Upper Virden highs (Fig. 11). Also it is apparent that at the time of latest Red Beds deposition, significant lows on the Upper Virden appeared immediately up-dip of the paleo-escarpment in both Routledge pools, particularly at Lsd. 4-24 and 12-24-9-25 W1 where the thickness exceeds 32 m (Fig. 11).

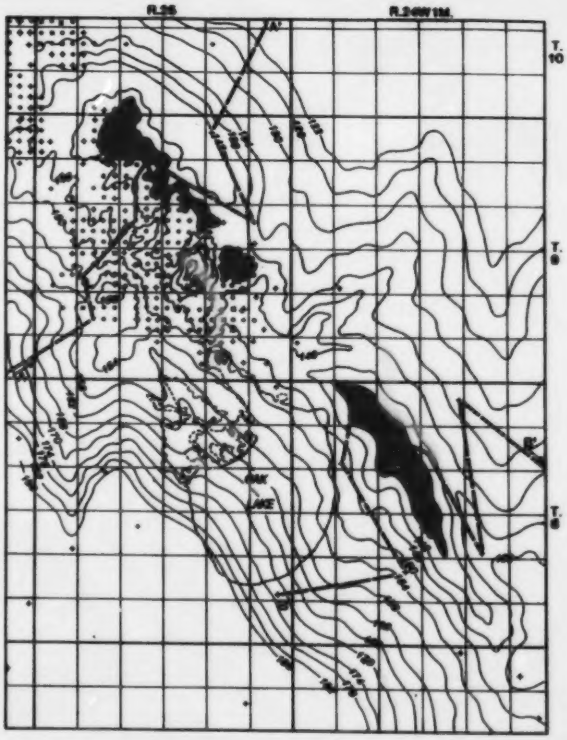

Figure 7: Structure contours on Upper Watrous shale marker, Routledge pool area; 4 m (13 ft) contour interval; stippled area less than -138 m (-453 ft) subsea elevation.

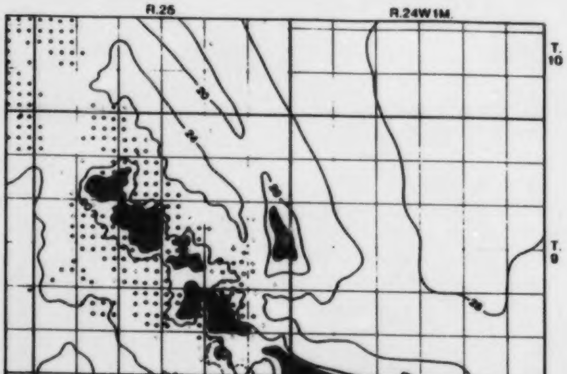

Figure 6: Isopach map, Lower Watrous Red Beds to Upper Watrous shale marker, Routledge pool area; 4 m (13 ft) contour interval; stippled area less than 18 m (60 ft) isopach.

Thinning of the combined Red Beds and Upper Virden strata within the Oak Lake area to less than 6 m indicates Upper Virden pseudo-structural highs on strike with those in the Routledge pools (Fig. 11). Of note again is the significant thinning on the structural closure at Lsd. 7-25-7-24 W1, similar to that in the East Routledge pool centred on Lsd. 8-10-9-25 W1. The isopach thick up-dip of the Routledge pools escarpment is not seen in the Oak Lake area, because the Virden Member and a portion of the Scallion Member are eroded within 1.5 km of the topographic high. However, Lsd. 3-14-8-23 W1, with 8.5 m (28 ft) of Red Beds and apparent thickening of the Upper Virden to Red Beds isopach *versus* Lsd. 5-27-8-24 W1 (Fig. 11), may in fact be situated near a companion low.

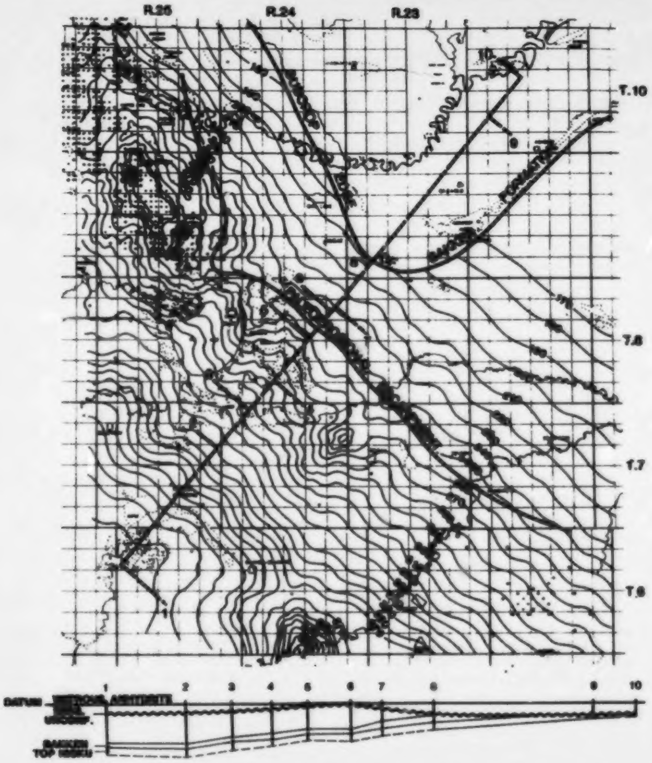

Figure 8: Structure contours on Upper Bakken Shale, Routledge pool area: 10 m (33 ft) contour interval; simplified structural dip section drawn on base Upper Watrous Anhydrite to Devonian Nisku Formation.

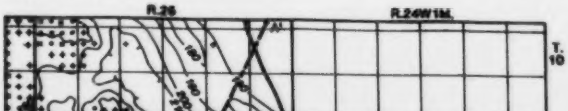

Figure 9: Structure contours on Scallion Member top, Lodgepole Formation, Routledge pool area: 10 m (33 ft) contour interval.

RELATIONSHIP OF STRUCTURAL ELEMENTS TO SALT DISSOLUTION TECTONICS

Structure in the Routledge pool area is believed to be in part related to dissolution of the Middle Devonian Prairie Evaporite salt overlying the Winnipegosis Formation. Trending roughly north across the Virden pool is a hinge separating a Winnipegosis eastern shelf from a western basin (Fig. 14). Pinnacle reefs up to 70 m (230 ft) in height are found throughout the basin and, as well, up to 115 m (377 ft) of Prairie halite. The latter thins to less than 55 m (180 ft) on the shelf. Throughout the Routledge area, and all but the western portion of the Virden pool, all of the Prairie salt has been removed by solution from off the shelf (Fig. 14).

Evidence of early salt solution is seen at Lsd. 7-25-7-24 W1 in the Oak Lake area in Bakken to Three Forks time, as indicated by the 26 m (85 ft) of thickening of these beds over regional values in an area of 1.5 to 3 km^2 (Fig. 12). Subsequent regional salt dissolution is recognized at the post-Mississippian pre-Watrous hiatus in surrounding wells by conversion of the 7-25 isopach thick into a closed structural high (Fig. 12). Timing of this structural inversion resulted in erosional thinning of subcropping Lodgepole Beds by up to 11 m (36 ft). Structural closure on Mississippian Beds is also seen within the Routledge pool area and is centred on Lsd. 8-10-9-25 W1 (Fig. 9) with up to 20 m (66 ft) of relief. These structural displacements resemble those in the Birdbear Formation at the Hummingbird pool in south-central Saskatchewan (Smith and Pullen, 1967) and the Nisku at the Wolf Creek nose, Roosevelt County, Montana (Swenson, 1967). Both are coincidentally positioned above the Winnipegosis shelf on the side of the Elk Point Basin opposite to the Routledge pool area.

A series of post-Mississippian collapse structures can be traced for 22 km (14 mi) along the Virden-Whitewater Lake trend through the Virden pool as a major, narrow (0.5 km/0.3 mi) northwesterly trending structural low. McCabe (1953) suggests that these structural lows are related to Prairie Salt dissolution controlled by basement structural elements or Winnipegosis reefs. In that structural relief on lows or highs does not vary considerably on either side of the salt dissolution edge, it seems more probable that synclinal lows are actually related to basement faulting. In the Routledge Pool area where no salt is preserved, the synclinal low seen up-dip of the escarpment is believed to be fault-controlled, with most movement having occurred at the post-Mississippian pre-Watrous unconformity time. Also, topographic closures on the Mississippian unconformity are most common in the no-salt area of the Virden-Routledge production trend, thereby suggesting some effects of multi-stage salt solution. Such closures coincide with Bakken and Three Forks thickening throughout most of the no-salt area (Martiniuk, 1988).

Finally, post-Watrous subsidence towards the Williston Basin centre has tilted all beds by up to 3 m/km (16 ft/mi), and so increased the structural relief of the subcropping reservoir beds.

Based upon the study of the Routledge pool area, a schematic cross-section drawn across the Virden-Whitewater ridge and the deep paleoerosional low to the northeast is presented in Figure 13. Clearly, underlying structural elements control the paleotopographic feature. Erosional backstepping appears to have advanced to the point where the Virden-Whitewater and Flossie Lake Members appeared as subcrops. Structural collapse up-dip and parallel to the paleo-escarpment resulted in preservation of a thicker Lodgepole section beneath the unconformity.

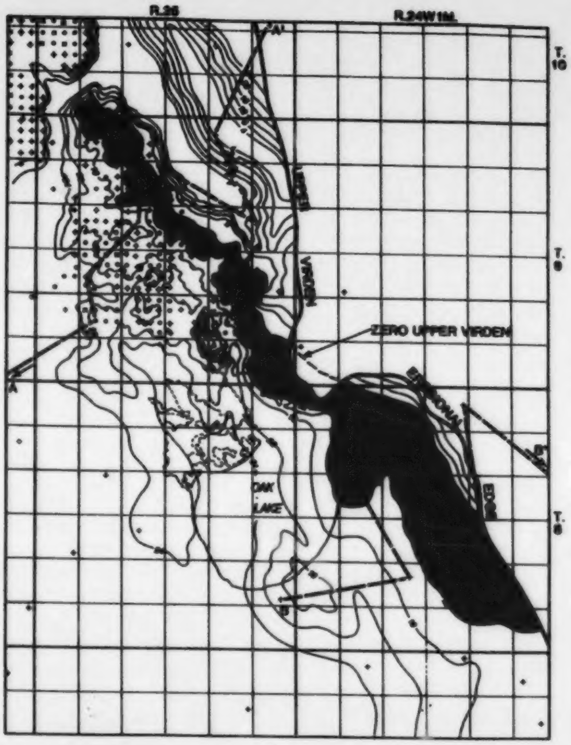

Figure 10: Isopach map, Upper Virden Member top to Mississippian unconformity, Routledge pool area; 4 m (13 ft) contour interval; stippled area less than 12 m (40 ft) isopach.

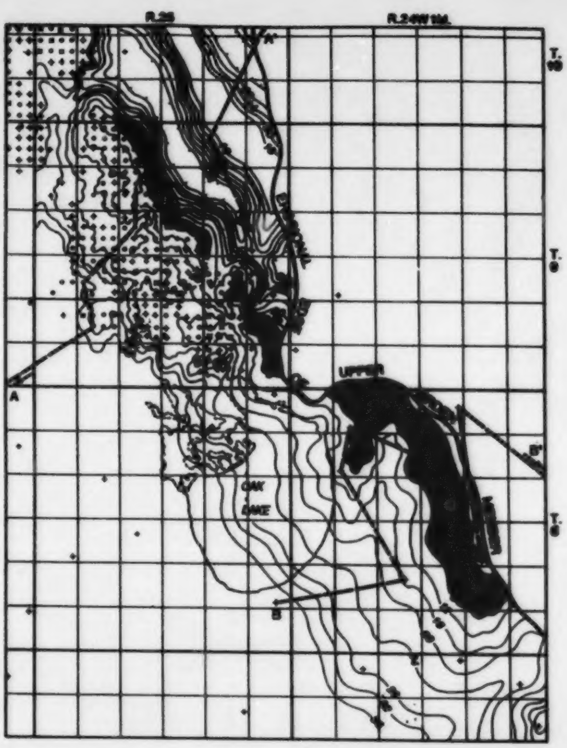

Figure 11: Isopach map, Upper Virden Member top to top Lower Watrous Red Beds, Routledge pool area; 4 m (13 ft) contour interval; stippled area less than 8 m (26 ft) isopach.

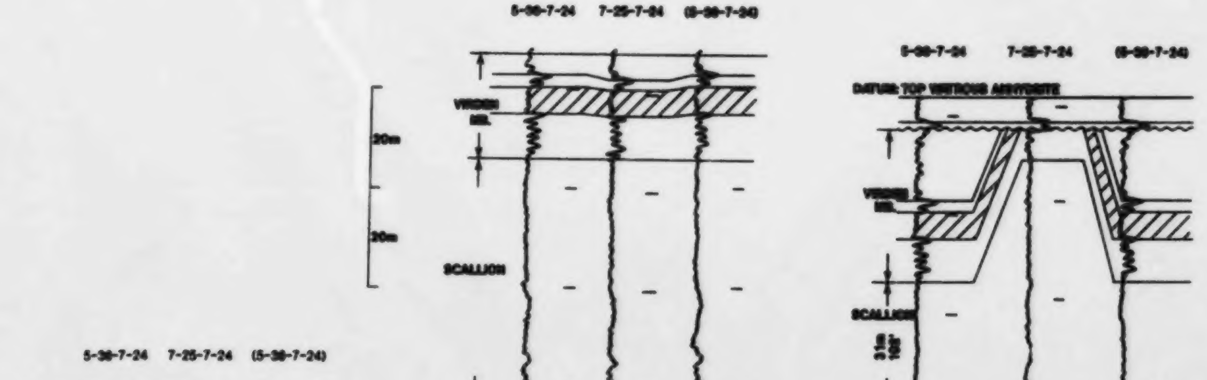

Figure 12: Multi-stage Prairie salt solution model, Oak Lake area. a) Three Forks time salt solution at 7-25-7-24 W1 by up to 27 m (89 ft); b) Deposition of Mississippian carbonate; c) Post-Mississippian salt solution at 5-36-7-24 W1 lowering Scallion Member by 31 m (102 ft) with Virden Member erosion upon unconformity high.

OIL TRAPPING IN THE ROUTLEDGE POOLS

Progressive erosional truncation of Lodgepole strata lowered the stratigraphic level of the producing interval in an up-dip direction across the Routledge pools (Fig. 4, Cross-section A-A'). Laterally, the oil was trapped against either northeast-trending structural lows or tight dolomite reservoirs. Subaerial exposure of Lodgepole strata at the Mississippian erosion surface resulted in extensive dolomitization and anhydrite-infilling of porosity, creating a 15 to 30 m (50 to 100 ft) thick impermeable caprock.

After the pattern of up-dip erosional truncation, the Whitewater Lake and Upper Virden reservoirs are oil productive nearest the southwestern edge of the Virden-Whitewater Lake ridge, followed in an up-dip direction by the Lower Virden and then the Scallion Member (Fig. 4). Each reservoir zone lies between -200 and -205 m (-656 to -673 ft) subsea and rises in an up-dip direction to the point where it becomes dolomitized as it nears the unconformity surface (Fig. 4, 5). Coincidentally, the producing level of the Upper Virden reservoir in the Souris Hartney pool (Fig. 1) lies at roughly -200 m (-656 ft) subsea as do most subcropping Lodgepole reservoirs between the two pools over a strike length of 35 km (22 mi).

In the Oak Lake area there are neither oil pools nor significant oil shows in spite of obvious examples of closures on subcropping reservoirs. One explanation for the lack of production may be the absence of a narrow structural low as seen up-dip of the escarpment in the Routledge pools (Fig. 4, 5).

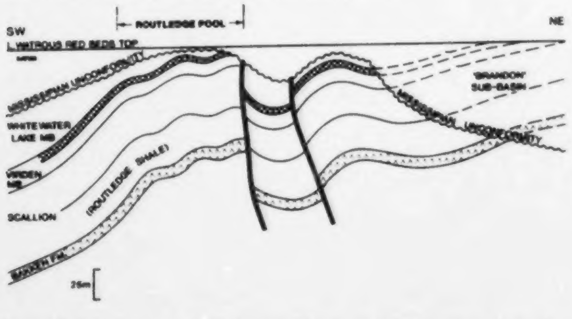

Figure 13: Schematic model: development of paleoerosional lineament, drawn from southwest to northeast across the Routledge pool and beyond to the northeast into the Brandon sub-basin; line of section is roughly 20 km (12 mi)

REGIONAL MODEL: DEVELOPMENT OF THE VIRDEN-WHITEWATER LAKE PALEOEROSIONAL TREND

Based upon geological relationships discussed in the Routledge pool area, a model for genesis of the Virden-Whitewater Lake trend is summarized as follows:

1. Nearest the northwestern end of the Virden-Whitewater Lake ridge is a southeast-plunging nose on the Bakken Formation (Birdtail Creek structural nose, Fig. 14). This nose is underlain by a complete section of Prairie Evaporite salt. Thus, the Birdtail structural nose is likely a basement-related structural element with parallel structural lows. This structural pattern presumably underlies the length of the trend across the no-salt area to at least the Souris-Hartney pool in Twp 6, Rge 22 W1. Northwesterly trending structural lows in the Virden-Routledge pools area may in part be related to late salt-dissolution.

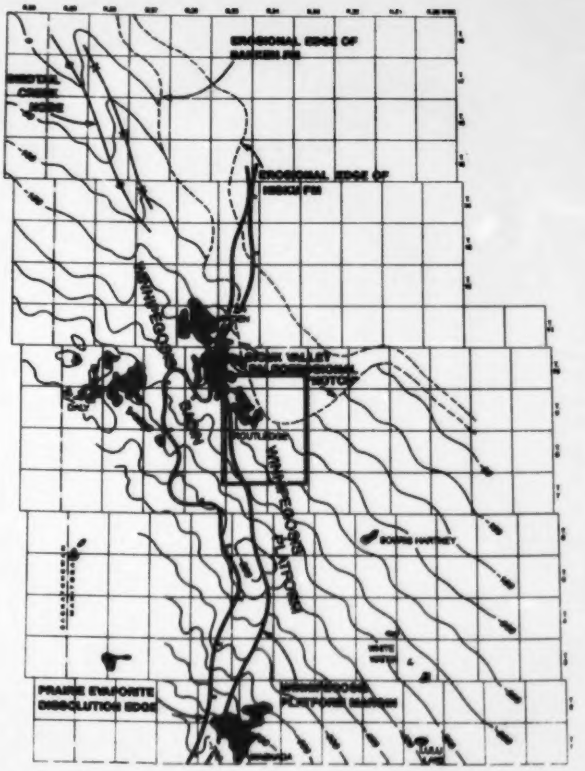
EROSIONAL EDGE OF BAKKEN FM
BIRDTAIL CREEK NOSE
EROSIONAL EDGE OF NISKU FM
WINNIPEGOSIS BASIN
SIOUX VALLEY PALEOEROSIONAL NOTCH
DALY
ROUTLEDGE
WINNIPEGOSIS PLATFORM
SOURIS HARTNEY
WHITE WATER
PRAIRIE EVAPORITE DISSOLUTION EDGE
WINNIPEGOSIS PLATFORM MARGIN

Figure 14: Regional structural contours on Upper Bakken Shale. 50 m (164 ft) contour interval; Winnipegosis Shelf margin and Prairie Evaporite salt edge indicated.

2. Coincident with the Birdtail structural nose is a post-Mississippian erosional high overlain by thin Watrous Red Beds and Anhydrite (Figs. 15, 16). As with Bakken structure, this topographic high plunges to the southeast to near the Souris Hartney pool (Twp 6; Rge 22 W1).

3. North of the Souris-Hartney Pool, the Lower Watrous Red Beds isopach map indicates a paleoerosional water divide separating to the southwest the Williston Basin and the Brandon sub-basin to the northeast (Fig. 16). Deep erosion of this sub-basin counter to regional dip down to at least the Nisku and Duperow Formations does not appear to coincide with any major structural elements.

   The Lodgepole ends in an escarpment capped by the Virden Member and overlooking the Brandon sub-basin. Deep erosional cuts perpendicular to the paleoridge are rarely seen; the Brandon sub-basin has no re-entrant across the ridge into the Williston Basin. The Sioux River paleoerosional notch (Fig. 16) apparently represents a major paleodrainage channel into the Brandon sub-basin. Northeast of the Souris-Hartney oil pool (Twp 6; Rge 22 W1) is the broad Scallion paleoerosional high with no capping Red Beds near the up-dip edge. Drainage up-dip from off the ridge to the northeast into the Sioux River notch and to the south into the main Williston Basin depocentre almost isolated this paleoerosional high as a large mesa.

4. South of Souris-Hartney all paleoerosional drainage appears to have been directed to the Williston Basin. A northerly trend of small isolated paleoerosional highs extends from at

least the international border (Figs. 15, 16). Oil is trapped at the Mississippian erosion surface in subcropping Whitewater Lake reservoir outliers (Martiniuk, 1986). In the east, drainage from the Brandon sub-basin eventually entered the Williston Basin, perhaps in Bottineau County, North Dakota.

5. Structure on the Mississippian unconformity along the paleoerosional axis rises from -300 m (-984 ft) in the Lulu Lake pool (Twp 1, Rge 21 W1) (Fig. 15) to -100 m (-328 ft) subsea on the Birdtail Creek structural nose in the north. The Virden and Routledge oil pools together represent one major oil accumulation draped across the paleoerosional high; at the Lower Watrous Red Beds depositional edge (Fig. 16) along the northwestern edge of the Virden pool, Lodgepole reservoirs beneath the unconformity are dolomitized, eliminating all permeability and porosity.

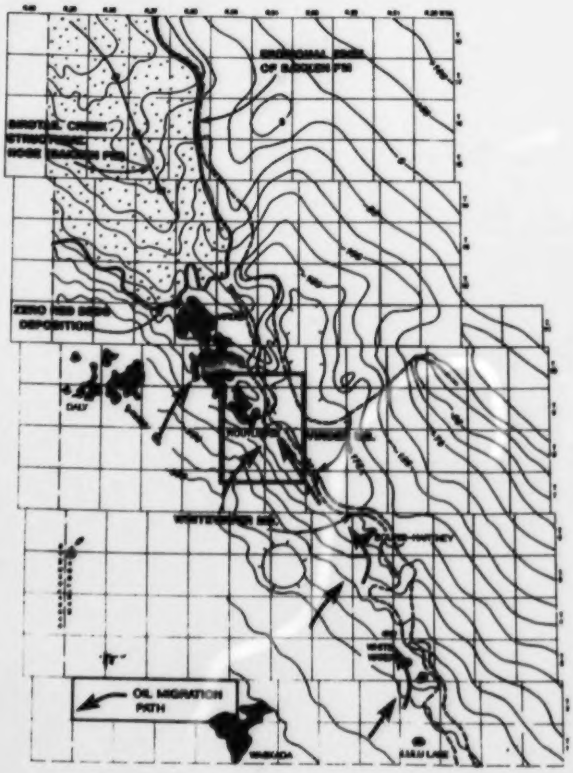

Figure 15: Regional structure contours on Mississippian unconformity along the Virden-Whitewater Lake trend; arrows indicate oil migration pathways.

ACKNOWLEDGEMENTS

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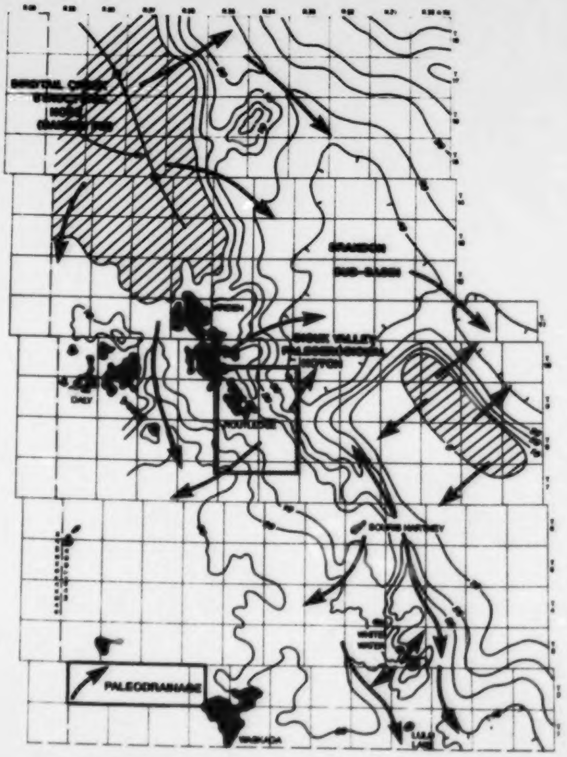

Figure 16: Regional isopach map, Lower Watrous Red Beds along the Virden-Whitewater Lake trend; 10 m (33 ft) contour interval; arrows indicate paleodrainage directions; zero Red Beds isopach stippled.

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STRATIGRAPHIC FRAMEWORK OF THE MISSISSIPPIAN LODGEPOLE FORMATION IN THE VIRDEN AND DALY OILFIELDS OF SOUTHWESTERN MANITOBA

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ABSTRACT

The Virden and Daly oilfields of southwestern Manitoba have produced more than 25 million m^3 (157 million barrels) of light (33-37° API) oil since their initial discovery. Limestone of the Early Mississippian (Kinderhookian) Lodgepole Formation is the principal reservoir rock; oil is structurally and stratigraphically trapped immediately downdip of the subcrop edge of this limestone.

Six stratigraphic units within the Lodgepole can be correlated across the study area. In the Virden area, a succession of bioclastic and oolitic grainstones and shales, recording shallow shelf deposition, onlap a prominent exposure surface capping a deeper water lime mudstone. To the west, in the Daly area, bioclastic limy shales are interbedded with thick packstone-wackestone accumulations, which represent some type of channel or mound feature, within a slope setting. Base-of-slope Waulsortian mud mounds have been described in eastern Saskatchewan immediately seaward of the shelf/slope package identified in this study. Detailed log correlations within the lower and middle part of the Lodgepole Formation reveal classic sigmoidal clinoform morphology which, coupled with the sedimentological data, suggests that the Lodgepole in the study area records the progradation of a westward-dipping carbonate ramp. The north-south orientation of the shelf/slope break in Manitoba is controlled by basement lineaments which have focused dissolution of Devonian salts and influenced subsidence rates during deposition of Lodgepole strata.

INTRODUCTION

In southwestern Manitoba, the Early Mississippian (Kinderhookian) Lodgepole Formation comprises a complex succession of shallow marine carbonates and shales which are the principal reservoir in the Virden and Daly oilfields (Figs. 1 and 2). These large oil accumulations, with combined produced and remaining reserves of more than 28 million m^3 (182 million barrels), are stratigraphically and structurally trapped near the subcrop edge of the Lodgepole Formation (Figs. 1 and 2). Vertical and lateral facies variations within the interval control the thickness and geometry of reservoir lithologies in both the Virden and Daly fields. Despite the economic significance of the Lodgepole interval in this part of the Williston Basin, stratigraphic correlations between these areas and the central part of the basin have never been definitely established, and the depositional history of these strata remains poorly understood. This integrated core and log study was initiated to resolve the stratigraphic correlation problems within the Lodgepole and to construct a depositional model, using concepts of carbonate stratigraphic sequence stratigraphy (Sarg, 1988), which might be used to explore for undiscovered oil accumulations along the northeastern rim of the Williston Basin.

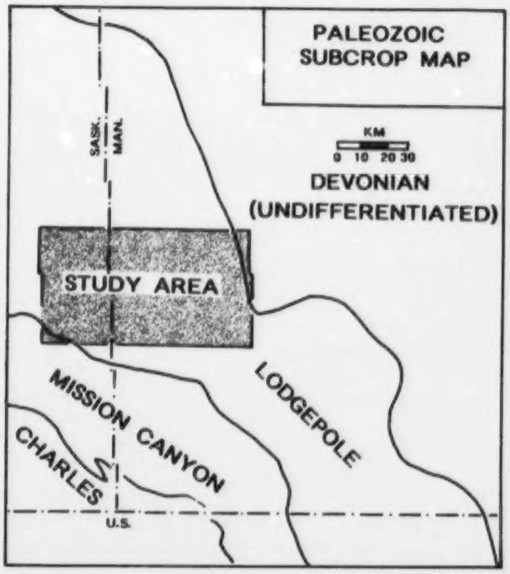

Figure 1: Location map showing study area and subcrop edges as defined by Kent (1984).

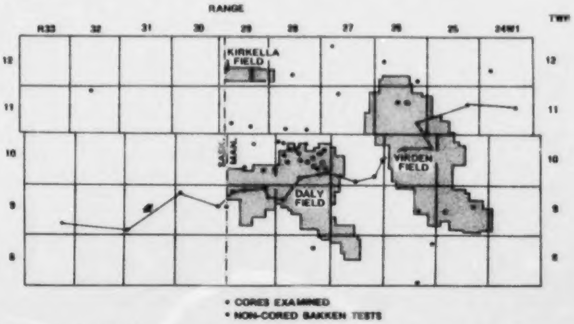

Figure 2: Detailed location map showing cores studied and cross-section described in this paper.

STRATIGRAPHY AND REGIONAL GEOLOGY

The Lodgepole Formation is an Early Mississippian (Kinderhookian) carbonate sequence which is present in most of the Williston Basin (Fig. 3). The term Lodgepole was introduced by Collier and Cathcart (1922) to describe the thinly bedded limestone and shale at the base of the Madison Limestone in the Little Rocky Mountains of northern Montana. In Manitoba, the Lodgepole Formation onlaps the Bakken Shale and is overlain by the carbonates and evaporites of the Mission Canyon Formation.

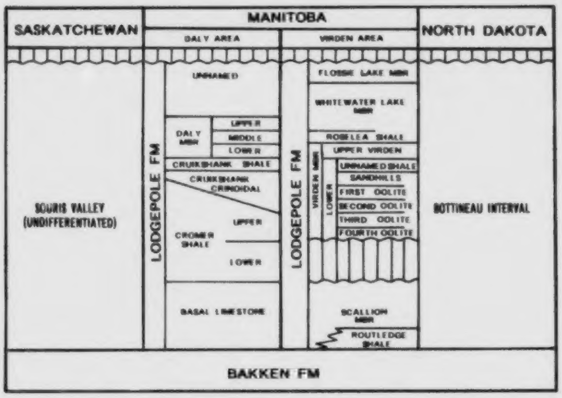

Figure 3: Stratigraphic column for Lower Mississippian strata along the northeastern margin of the Williston Basin.

The Lodgepole reaches a maximum thickness of more than 150 m in southwestern Manitoba (McCabe, 1959) and thins to zero at its subcrop edge where it is overlain by Jura-Triassic redbeds of the Amaranth Formation. In the North Dakota and Saskatchewan parts of the Williston Basin, the Lodgepole-equivalent strata (Bottineau Interval and Souris Valley Beds respectively) comprise a monotonous (undivided) section of argillaceous limestones and shales (Fig. 3). However, near their subcrop edge in Manitoba, these strata are made up of complexly interbedded shales, mudstones, and grain-supported limestones and dolomites. Previous workers (Ower, 1953; Milne and Nickoloff, 1956; Berg, 1956; Organ and Russin, 1956; Stanton, 1955, 1958; McCabe, 1959, 1963) constructed different nomenclature schemes for the Virden and Daly areas.

In the Virden area, the Lodgepole is divided into the Scallion, Virden, Whitewater Lake and Flossie Lake Members. A black shale locally developed at the base of the Scallion is termed the Routledge facies of the Scallion Member (Fig. 3). The Virden Member is subdivided into Upper and Lower units and the Lower Virden is, in turn, subdivided into five oolitic limestone/shale cycles which are termed, in descending order, the Sandhill, First, Second, Third and Fourth Oolite Beds, (Stanton, 1955; McCabe, 1959, 1963). In the Daly area, the Lodgepole interval is subdivided into the Basal Limestone, Cromer Shale, Cruickshank Crinoidal, Cruickshank Shale, Daly Member, and an upper Unnamed Member (Organ and Russin, 1956; McCabe, 1959).

PREVIOUS STUDIES

Most papers dealing with the Lodgepole carbonates of southwestern Manitoba were published during the development of the field in the late 1950s (Berg, 1956; Stanton, 1955, 1958; Organ and Russin, 1956). A major synthesis of the stratigraphy, petrography and petroleum gelogy of these strata was published by McCabe (1959, 1963) and a detailed discussion of the diagenetic history was published by Young and Greggs (1975).

DATA BASE

Approximately 1 700 m of core were examined from 34 wells in a 200 km^2 area encompassing the Virden and Daly oilfields (Fig. 2). Most of the cores examined were from wells which penetrate the entire Lodgepole, such that the top of the Bakken could be used as a stratigraphic datum on cross sections. In the eastern part of the study area, an excellent core data base exists as at least part of the Lodgepole is cored in most wells and the entire interval is cored in several wells. However, in eastern Saskatchewan, only a few wells penetrate the entire Lodgepole interval and few are cored. As a result, the data base is poor in this area (Fig. 2). The cores were described using Dunham's (1962) terminology (Table 1) and the core data were tied into a regional east-west stratigraphic cross section (Fig. 8) to resolve the stratigraphic correlation problems within the Lodgepole interval.

Table 1: Depositional facies recognized in this study.

1) Lime/Dolomite Mudstone	A	Clean (non argillaceous) lime mudstone
	B	Interbedded dolomite mudstone/anhydrite
	C	Limy bioclastic shale - argillaceous lime mudstone
	D	Siliceous lime mudstone with 10-30% white chert nodules
2) Wackestone		Bioclastic non argillaceous wackstone
3) Packstone/Grainstone	A	Bioclastic (crinoid-coral-brachipod) grainstone
	B	Oolitic bioclastic (crinoidal) grainstone
	C	Intraclastic grainstone/breccia

DEPOSITIONAL FACIES

Three depositional facies are recognized: lime or dolomite mudstones (Facies 1), wackestones (Facies 2), and grainstone/packstones (Facies 3). Facies 1 and 3 are further subdivided into seven subfacies on the basis of colour, shale content and mineralogy. These subfacies are listed in Table 1.

LIME OR DOLOMITE MUDSTONES

Mud-supported carbonates (Facies 1) make up the bulk of the Lodgepole interval in the study area and do not constitute important reservoir lithologies.

Subfacies 1A consists of pink, light grey and cream-coloured, non-argillaceous lime mudstones. It contains rare solitary corals and crinoids. *Zoophycos, Helminthopsis* and *Planolites* trace fossils were noted in a few cores (Fig. 5B). Overall, the unit is massive to crudely stratified. This subfacies is restricted to the Basal Limestone in the Daly area and the stratigraphically equivalent Scallion Member in the Virden area.

Subfacies 1B is a light grey to pink non-argillaceous dolomite mudstone which contains up to 20 percent bands and blebs of blue-grey anhydrite and up to 10 percent bands of silicified limestone. The subfacies is restricted to the upper few metres of the Lodgepole (Unnamed Member) in the Daly-Ebor area. The proportion of anhydrite, and degree of silicification both increase towards the pre-Amaranth unconformity, suggesting that this is a diagenetic mineralogy-texture associated with the overlying exposure surface. Subfacies 1B is typically oil-stained although it exhibits only minor vuggy or solution-enhanced fracture porosity and little or no matrix porosity.

Subfacies 1C is a highly argillaceous lime or dolomite mudstone which grades into limy shale (Fig. 5C). It has a mottled greenish-grey to maroon coloration and contains abundant crinoid, brachiopod, coral and fenestellid bryozoan bioclasts floating in a very argillaceous matrix. The subfacies contains up to 10 percent sharp-based grainstone-packstone beds up to 10 cm in thickness. This subfacies comprises most of the Cromer Shale interval in the Daly-Kirkella areas and the "inter- oolite" beds of the Lower Virden Member in the Virden area.

Subfacies 1D comprises light grey, pink or white, non-argillaceous lime mudstone which contains 5 to 30 percent amorphous blebs or nodules of white tripolitic chert (Fig. 5D). It constitutes most of the Scallion and Basal Limestone Members in the Virden and Daly areas respectively. Although tight in most parts of the study area, it is an excellent reservoir in the northern part of the Virden pool where localized leaching and dissolution of the carbonate mud fraction has created very high porosity and permeability (Berg, 1956).

WACKESTONES

Facies 2 consists of tan to light brown, non-argillaceous lime wackestones which contains abundant crinoid and brachiopod bioclasts. It is restricted to the Lower Daly and Cruickshank Crinoidal Members of the Daly area and the Upper Virden Member in the Virden pool; a particularly thick (13.5 m) section is found in the 14-4-10-28 W1 well (Fig. 6) within the Cruickshank Crinoidal Member. This facies generally exhibits low to moderate porosity and permeability.

GRAINSTONES/PACKSTONES

Facies 3 consists of clean (non-argillaceous) lime grainstone and packstones which form the principal reservoir in the Virden and Daly fields. It is subdivided into three subfacies based on the proportion and type of grainy material present.

Subfacies 3A consists of pink, light grey to brown bioclastic lime grainstones and packstones which contain abundant crinoid, brachiopod and solitary coral fragments (Fig. 4B). It is crudely bedded and exhibits low angle (<10°) laminations. No high angle cross-bedding was noted. The low angle convergent laminations may represent hummocky cross-stratification. This subfacies makes up most of the Upper Virden Member in the Virden area and the Cruickshank Crinoidal and Lower Daly in the Daly and Kirkella pools. It exhibits poor to fair porosity and medium brown oil staining within the main Virden and Daly pools and is tightly cemented and non-stained in the structurally low areas between the fields.

Subfacies 3B includes buff, light grey and brown oolitic grainstones and packstones (Fig. 4A) which are generally structureless but occasionally exhibit low angle convergent lamination similar to that described in subfacies 3A. No high angle cross-bedding was noted. The oolitic subfacies is present only in the Lower Virden Member in the eastern party of the study area. The oolitic limestones exhibit poor to fair porosity and light brown staining in the Virden pool but are tightly cemented in the structurally low areas between the pools.

Subfacies 3C consists of white to light grey, coarse grainstones and/or intraclastic breccias which contain angular clasts of lime mudstone in a bioclastic and/or oolitic matrix (Fig. 5A). This subfacies typically occurs as a thin (<2 m) lag at the base of two fining-upward successions at the base of the Lower Virden Member in the eastern part of the study area.

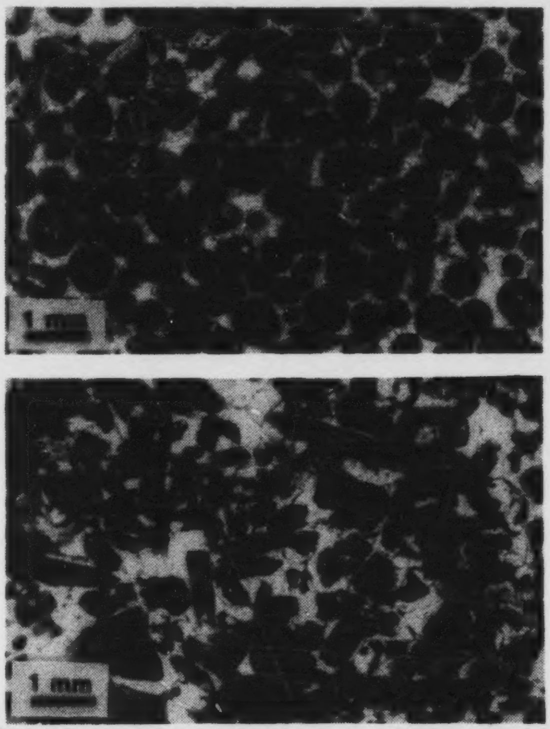

Figure 4: Photomicrographs of (a) Subfacies 3B, oolitic grainstone, 4-35-8-25 W1, depth 2136' (651 m) and (b) Subfacies 3A, bioclastic grainstone, 13-21-12-28 W1, depth 2188'(666.9 m).

INTERPRETATION OF DEPOSITIONAL FACIES

The entire Lodgepole interval appears to have been deposited under subtidal, open marine conditions across a shallow, westward-dipping ramp. The presence of an open-marine fauna (crinoids, brachiopods, bryozoans, and solitary and colonial corals) suggests that the bottom waters were well oxygenated and of normal salinity. At least two unconformities are recognized in the Lodgepole section although no evidence has been found of inter- to supratidal conditions (*e.g.*, mudcracks, stromatolites, *etc.*) during deposition of the Lodgepole interval within the study area.

More detailed interpretations of the various depositional facies are more difficult to make as most facies are devoid of any depth-diagnostic sedimentary structures. In general, the mud-supported lithologies are probably indicative of a lower energy (deeper water?) depositional setting whereas the grain-supported lithologies represent a shoaling environment in which wave action winnowed the mud fraction from the grainy components. The trace fossils noted in the lime mudstones (Facies 1A) including *Zoophycos*, *Planolites*, *Helminthoides* and *Helminthopsis* are characteristic of relatively deep, low energy marine environments.

Within the bioclastic and oolitic grainstones, the absence of high angle cross bedding is enigmatic since most modern shelf shoals are dominated by trough or planar-tabular cross bedding (Halley *et al.*, 1983). By analogy with clastic systems, the absence of high angle cross-stratification and abundance of low angle convergent lamination (interpreted as hummocky cross-stratification) may indicate deposition below fairweather wave base, in a lower shoreface or shelf setting (Walker, 1984).

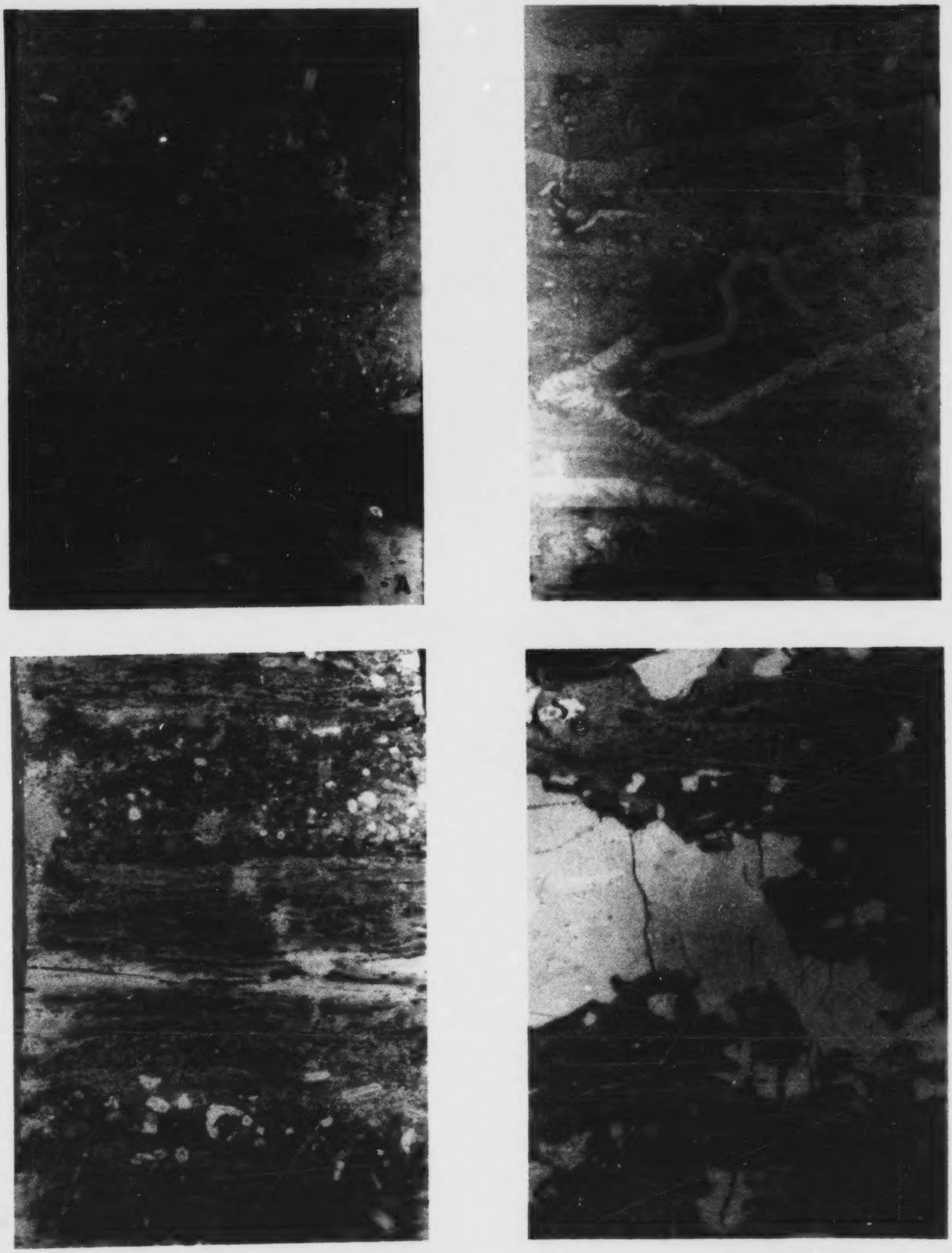

Figure 5: Core photographs showing (a) intraclast-bearing grainstone (Subfacies 3C, Lower Virden) overlying reddish paleosol (?); (b) non-argillaceous lime mudstone (Subfacies 1A, Unnamed Member) with *Zoophycos* burrows; (c) argillaceous lime mudstone (Subfacies 1C, Cromer Shale), containing fenestellid bryozoan debris and sharp-based crinoidal grainstone beds; (d) non-argillaceous lime mudstone with irregular tripolitic chert nodules (subfacies 1D, Scallion Member). Bar scale = 1 cm.

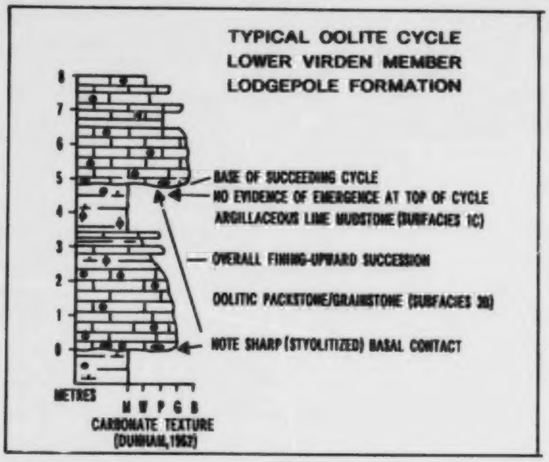
TYPICAL OOLITE CYCLE
LOWER VIRDEN MEMBER
LODGEPOLE FORMATION
8
7
6
5
4
3
2
1
0
BASE OF SUCCEEDING CYCLE
NO EVIDENCE OF EMERGENCE AT TOP OF CYCLE
ARGILLACEOUS LIME MUDSTONE (SUBFACIES 1C)
OVERALL FINING-UPWARD SUCCESSION
OOLITIC PACKSTONE/GRAINSTONE (SUBFACIES 3B)
NOTE SHARP (STYOLITIZED) BASAL CONTACT
METRES
M W P G B
CARBONATE TEXTURE
(DUNHAM, 1962)

Figure 6: Idealized description of a Lower Virden oolitic cycle within Virden field.

The abundance of current-baffling organisms such as fenestellid bryozoans and crinoids in subfacies 1C in the Cromer and Cruickshank Member in the Daly area may represent some type of base-of-slope environment. Similar facies have been described in Lodgepole-equivalent strata elsewhere in North America (Smith, 1982; Stone, 1972; MacQuown and Perkins, 1982; King, 1986; Sereda and Kent; 1987, Ausch and Meyer 1990).

The variation in the clastic component within the Lodgepole probably reflects fluctuations in relative sea level within the basin. The high argillaceous content of subfacies 1C may record deposition during a relative low stand in sea level when more terrigenous material was being delivered to the basin (Wilson, 1975). Non-argillaceous lime mudstones of the Scallion-Basal Limestone and Whitewater Lake Members may record deposition during a high stand in sea level during which only minor amounts of terrigenous clays were introduced to the basin.

FACIES SEQUENCES IN THE LODGEPOLE FORMATION - VIRDEN AREA

In this study, the subdivision of the Lodgepole employed by Stanton (1955) is used although the contacts of several of the units have been modified slightly. The basal Scallion Member comprises a thick monotonous succession of clean (non-argillaceous) cherty lime mudstones (Subfacies 1D). The basal contact with the radioactive Bakken shales is abrupt and the upper contact is marked at the base of the intraclastic breccia/grainstones (Fig. 5A) developed at the base of the Lower Virden, immediately below the fourth oolite (*e.g.*, 5-1-12-26 W1, Fig. 9). Locally, the top few metres of the Scallion lime mudstones are dolomitized, have a reddish coloration, and are interpreted as "terra rosa" paleosol (Fig. 9). The intraclastic grainstones capping the cherty mudstones are included in the Scallion interval by most workers. However, these strata are more similar to the overlying "grainy" sequence in the Lower Virden and thus included in that interval in this study.

The overlying Lower Virden Member consists of a series of thin, sharp-based oolitic and/or bioclastic beds (Subfacies 3A, B), interbedded with argillaceous lime mudstones (Subfacies 1C). Each of the five cycles within the Lower Virden has an abrupt base (accentuated by stylolitization) in which a centimetre-thick basal bioclastic grainstone is capped by oolitic grainstones which grade upwards into argillaceous lime mudstones (Fig. 6). All five oolite beds are recognized in only a few cores (*e.g.*, 13-17-9-25 W1), and in most wells, one or more of the oolite beds has pinched out into the surrounding argillaceous shale or undergo a lateral (westward) facies change into thin bioclastic grainstones. The distribution of individual oolite beds is difficult to map with the limited core control outside field boundaries although the oolitic beds clearly extend well beyond the boundaries of the Virden Pool.

The Upper Virden consists of a thick (2.0 to 6.0 m) bioclastic grainstone/packstone (Subfacies 3A) which is overlain by reddish maroon shale (Subfacies 1C) of the Roselea Marker. The Upper Virden and Roselea are laterally continuous and can be correlated throughout the eastern part of the study area. The Whitewater Lake and Flossie Lake Members do not comprise important reservoirs in the study area and were not examined in this study.

INTERPRETATION OF THE VIRDEN SUCCESSION

The thick succession of clean, cherty lime mudstones comprising the Scallion Member record deposition in a relatively low energy ramp environment with minimal clastic influx. The rapid basinward-thinning of this carbonate mud-rich succession is similar to "catch-up" platform sequences which are considered by Sarg (1988) to record slow sediment accumulation rates. The grain-supported lithologies (Facies 3) noted in one core within the Scallion (*i.e.* 5-3-10-27 W1) may reflect winnowing along a narrow shelf crest bordering a muddy carbonate platform. The widespread distribution of the two intraclastic zones at the top of the Scallion suggests that they represent exposure (disconformity?) surfaces. This interpretation is supported by the recognition of a brick-red alteration zone (terra rosa paleosol?) at the base of the lower unit in the 5-11-12-26 W1 core (Fig. 9). In several cores (*e.g.* 5-11-10-26 W1), the degree of leaching and porosity development and the proportion of white tripolitic chert nodules and dolomitization are highest immediately below this breccia interval suggesting that these features are somehow related to this (Mississippian-aged) exposure event. The high degree of leaching at this surface, as described by Berg (1956), has involved dissolution of both the tripolotic chert and carbonate phases, suggesting that leaching and porosity development postdated the silicification.

The overlying Lower Virden comprises interbedded oolitic and bioclastic grainstones and argillaceous mudstone which probably record eustatic sea level fluctuations across a broad shallow shelf. The sharp base of each of the oolitic horizons (Fig. 6) may represent a sequence boundary, analogous to those recently described in Pleistocene carbonates in the Barbados by Humphrey and Kimbell (1990). These surfaces record a drop in relative sea level which may have led to exposure of the shallow shelf. The oolitic lenses capping each sequence boundary were probably deposited during the initial stage of the subsequent transgression. The gradational contact of the oolitic grainstones and the overlying argillaceous shale (Fig. 6) suggests that continued deepening eventually led to drowning of the oolite shoal complex. This interpretation implies that the Lower Virden oolitic sequences record only the transgressive phase of the cycle and that the only evidence of subsequent regression is the sharp base of the succeeding sequence. This process appears to have repeated itself at least five times, resulting in the complex cyclicity recognized in the Lower Virden interval.

The thicker bioclastic grainstones and packstones in the Upper Virden also probably indicate a widespread shoaling of the basin.

The absence of oolites suggests that the basin did not shoal to the same extent as it did during deposition of the Lower Virden. The overlying shales of the Roselea Marker apparently record a major transgression as they underlie lime mudstones (Lower Whitewater Lake Member) which appear to imply a return to low energy, deeper water conditions.

FACIES SEQUENCES IN THE LODGEPOLE FORMATION - DALY AND KIRKELLA AREAS

Six stratigraphic units can be recognized in the Daly and Kirkella areas and these include, in ascending order, the Basal Limestone, Cromer Shale, Cruickshank Crinoidal, Cruickshank Shale, Daly, and Unnamed Members. In addition, the Daly Member is subdivided into Lower, Middle and Upper units (Organ and Russin, 1956; McCabe, 1959, 1963). Two wells (14-4-10-28 W1 and 10-12-10-28 W1) in the Daly area which have cored the entire Lodgepole interval and show thickness variations between the various stratigraphic units are shown in Figures 7 and 8.

The Basal Limestone consists of a westward-thinning wedge of cherty, lime mudstones (Subfacies 1D) which are texturally very similar to the thicker Scallion interval in the Virden area.

The overlying Cromer Shale consists of argillaceous lime mudstones and shale (Subfacies 1C) which exhibit significant thickness variations across only a few kilometres. The Cromer

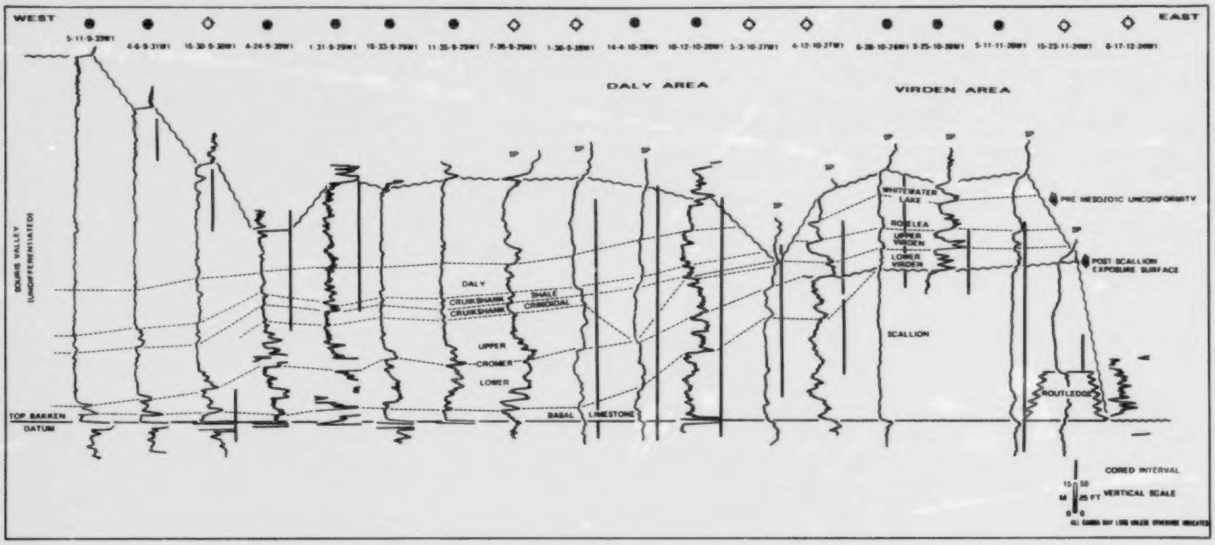

Figure 7: Stratigraphic cross-section (top of Bakken as datum) using gamma-SP log traces.

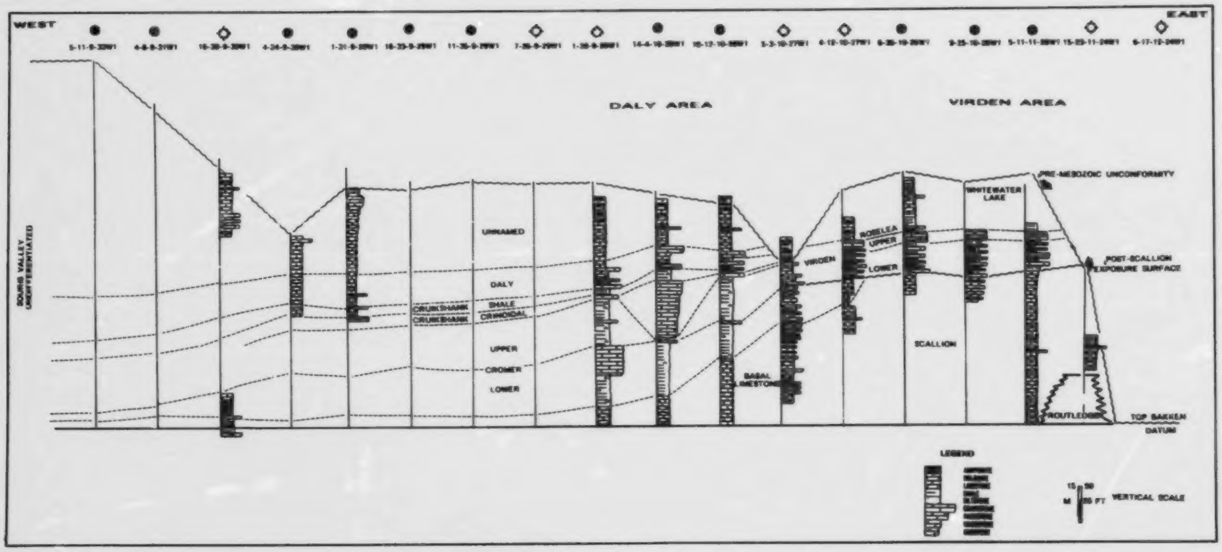

Figure 8: Stratigraphic cross-section (top of Bakken as datum) using core data.

reaches a maximum thickness of 45 m in 10-12-10-28 W1 and thins to only 21 m in 14-4-10-28 W1. The thinning of the Cromer is compensated by a thickening of the overlying Cruickshank Crinoidal, but whether this results from erosional channelling of the underlying Cromer Shale or is due to a lateral facies from limy shales into some type of grain-supported mound or buildup within the lower part of type Cruickshank Crinoidal is unclear. The anomalously thick Cruickshank Crinoidal unit in the 14-4-10-28 W1 well has a sharp scoured base and is completely devoid of shale partings, suggesting that it is some type of younger channel-fill feature which has been incised down into the older Cromer Shale (*i.e.* it is not a lateral facies-equivalent of these shales). Since the Cruickshank Crinoidal-Cromer contact is cored in only a few wells and is not even penetrated in many wells, the thickness, geometry and origin of the Cruickshank Crinoidal interval remain unclear.

The Cruickshank Shale is a thin shaly limestone (Subfacies 1C) which blankets the entire Daly area. The overlying Lower Daly consists of a series of bioclastic grainstones (Subfacies 3A) interbedded with argillaceous shales (Subfacies 1C). Siliceous lime mudstones (Subfacies 1D) are also locally developed in the Lower Daly. The Middle Daly is characterized by a distinctive clean lime mudstone which contains up to 25 percent white chert nodules and is capped by a thin, reddish crinoidal packstone. A similar facies sequence was noted in the basal part of the Whitewater Lake Member in the Virden area, suggesting that these are stratigraphic equivalents.

The upper part of the Lodgepole interval, referred to as the Unnamed Member by Organ and Russin (1956), consists of clean carbonate mudstone (Subfacies 1A, 1B) across most of the Daly-Kirkella area. In some cores (*e.g.*, 1-31-10-29 W1), a thin bioclastic grainstone is developed near the top of this interval. In general, the Unnamed Member is dolomitized and partially replaced by anhydrite, particularly adjacent to the Pre-Jurassic unconformity.

Most of the oil production in Daly and Kirkella areas is derived from porous bioclastic grainstones in the Cruickshank Crinoidal and Lower Daly intervals. The pay ranges in thickness up to 10 m but averages 3 to 5 m in the Daly field and less than 2 m in the Kirkella field (Fig. 2). Minor hydrocarbon production is also derived from porous grainstones (*e.g.*, 1-31-10-29 W1) and from dolomitized mudstones at the top of the Unnamed Member.

SASKATCHEWAN SUCCESSION

Near the Manitoba/Saskatchewan border, the geophysical log character of the Lodgepole (Souris Valley) succession changes dramatically. The Lodgepole is cored in only a few wells as it is not an important reservoir in this area. However, the log signature indicates that the Lodgepole comprises a relatively monotonous succession of slightly argillaceous, non-porous carbonates which overlie a thin (<5 m) shale developed near the base of the Lodgepole. Waulsortian mud mounds (core and flank facies) have been recognized from the top of the Lodgepole in this area by Sereda and Kent (1987).

STRATIGRAPHIC CORRELATIONS BETWEEN THE VIRDEN, DALY AND KIRKELLA AREAS

The correlation of six stratigraphic units within the Lodgepole is illustrated in Figures 7 and 8. The salient features of the correlations are:

1) The Scallion (Virden area) and Basal Limestone (Daly area) are approximate stratigraphic equivalents. These strata record an episode of relatively deep (below wave base) subtidal sedimentation across a gently westward-dipping ramp. This interval thins markedly to the west and, in eastern Saskatchewan, is represented by a very thin (< 5 m) limestone underlying the shale marker described above (Figures 7, 8). The grain-supported lithologies noted in the west Virden area (*i.e.*, 5-3-10-27 W1) may represented a narrow shelf crest facies fringing a shallow shelf deposit (Fig. 11A).

2) Eastwards, the argillaceous mudstones of the lower Cromer Member of the Daly area either pinch out or change facies as no shaly facies are found at the same stratigraphic interval in the Virden area. The basal part of the Cromer shales probably records deposition during a low stand in sea level (Fig. 11B) at which time the upper part of the Scallion Member of the Virden area would have been subaerially exposed. The concentration of current-baffling organisms within the Cromer Shale, such as fenestellid bryozoans and crinoids, suggests that these organisms may have focused deposition of terrigenous-derived material in a base-of-slope setting (Fig. 10), analogous to lithologies described in the late Mississippian strata in Kentucky (Ausch and Meyer, 1990). Farther basinward (west), strata equivalent to the Cromer (Fig. 7) contain a much lower argillaceous component. The grain-supported lithologies at the top of the Lower Cromer record a minor shoaling event, preceding deposition of the upper Cromer Shale.

3) The upper Cromer Shale is correlated with the interbedded oolitic grainstones and shaly limestones of the Lower Virden Member. The oolitic beds apparently pinch out to the west (Figs. 7, 8) and the intervening argillaceous mudstones merge to form the upper part of the Cromer Shale in the Daly area. The cyclicity recognized in the Lower Virden probably records recurrent fluctuations in relative sea level in a shallow shelf setting. Correlation lines in this part of the succession exhibit classic sigmoidal clinoform patterns which are considered characteristic of a prograding shelf/slope complex (Vail *et al.*, 1977; Sarg, 1988; Humphrey and Kimbell, 1990).

4) The Cruickshank Crinoidal and Cruickshank Shale Members of the Daly area respectively correlate with the uppermost oolitic cycle (Sandhill Oolite) and the unnamed shale capping the Lower Virden Member in the Virden area. The thick packstone/ wackestone accumulations within the Cruickshank Crinoidal in the Daly area (*e.g.*, 14-4-10-28 W1) are enigmatic in origin but probably represent some type of channel-fill succession (Fig. 11D). This may be related to a marked drop in relative sea level which caused the channels to incise into the underlying Cromer Shales. The ensuing transgression deposited wackestones and packstones of the Cruickshank Crinoidal in the Daly area and the Sandhill oolite beds across the shelf to the east. The peak of this transgression is marked by the Cruickshank Shale in the Daly area and the lithologically similar (unnamed) shale separating the Sandhill Oolite from the Upper Virden grainstones in the Virden area.

5) The bioclastic grainstones of the Lower Daly Member correlate with similar lithologies in the Upper Virden Member in the Virden pool (Fig. 11E). This unit forms a relatively thin, sheet-like deposit, and records a widespread shoaling event across a broad shallow shelf. It overlies the clinoform shelf/slope succession (Cromer- Cruickshank interval).

6) The cherty lime mudstones of the Middle Daly Member in the type area are correlated with similar lithologies at the base of the Whitewater Lake Member in the Virden area. This

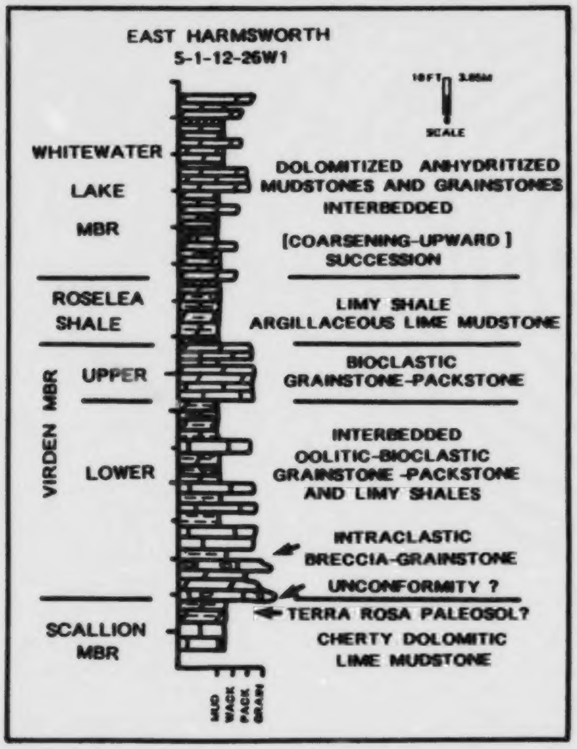
EAST HARMSWORTH
5-1-12-26W1
10FT 3.05M
SCALE
WHITEWATER LAKE MBR
DOLOMITIZED ANHYDRITIZED MUDSTONES AND GRAINSTONES INTERBEDDED
[COARSENING-UPWARD] SUCCESSION
ROSELEA SHALE
LIMY SHALE ARGILLACEOUS LIME MUDSTONE
VIRDEN MBR
UPPER
BIOCLASTIC GRAINSTONE-PACKSTONE
LOWER
INTERBEDDED OOLITIC-BIOCLASTIC GRAINSTONE -PACKSTONE AND LIMY SHALES
INTRACLASTIC BRECCIA-GRAINSTONE
UNCONFORMITY ?
TERRA ROSA PALEOSOL?
SCALLION MBR
CHERTY DOLOMITIC LIME MUDSTONE
MUD
WACK
PACK
GRAIN

Figure 9: Typical core description from the Virden area pool (5-1-12-26 W1) showing two terra rosa paleosol horizons capping a Mississippian-age exposure surface developed above Scallion Member, overlain by interbedded limy shales and oolitic bioclastic grainstones and packstones of the Virden Member.

represents the base of another major depositional cycle (Fig. 11F). A thin sharp-based crinoidal grainstone which caps the Middle Daly can be recognized across the study area and probably records a minor drop in relative sea level. The clean carbonate mudstones of the upper Unnamed Member of the Lodgepole are probably equivalent to the Whitewater Lake Member (Stanton, 1955) in the Virden area.

EXPLORATION IMPLICATIONS AND RECOMMENDATIONS

The Virden, Daly and Kirkella fields are complex stratigraphic, structural and diagenetic hydrocarbon traps. These fields differ from the Mississippian subcrop traps in other parts of the Williston Basin, the distribution of which is more directly

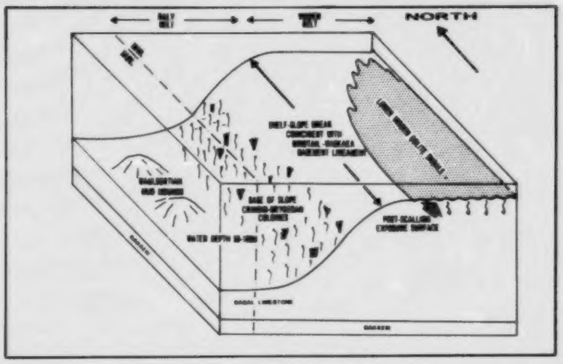
NORTH
BASE OF SLOPE CRINOID-BRYOZOAN COLONIES
POST-SCALLION EXPOSURE SURFACE
BASAL LIMESTONE
BAKKEN

Figure 10: Schematic depositional model showing locations of the shelf/slope break and bryozoan-crinoid colonies which trapped terrigenous-derived muds at the base of slope.

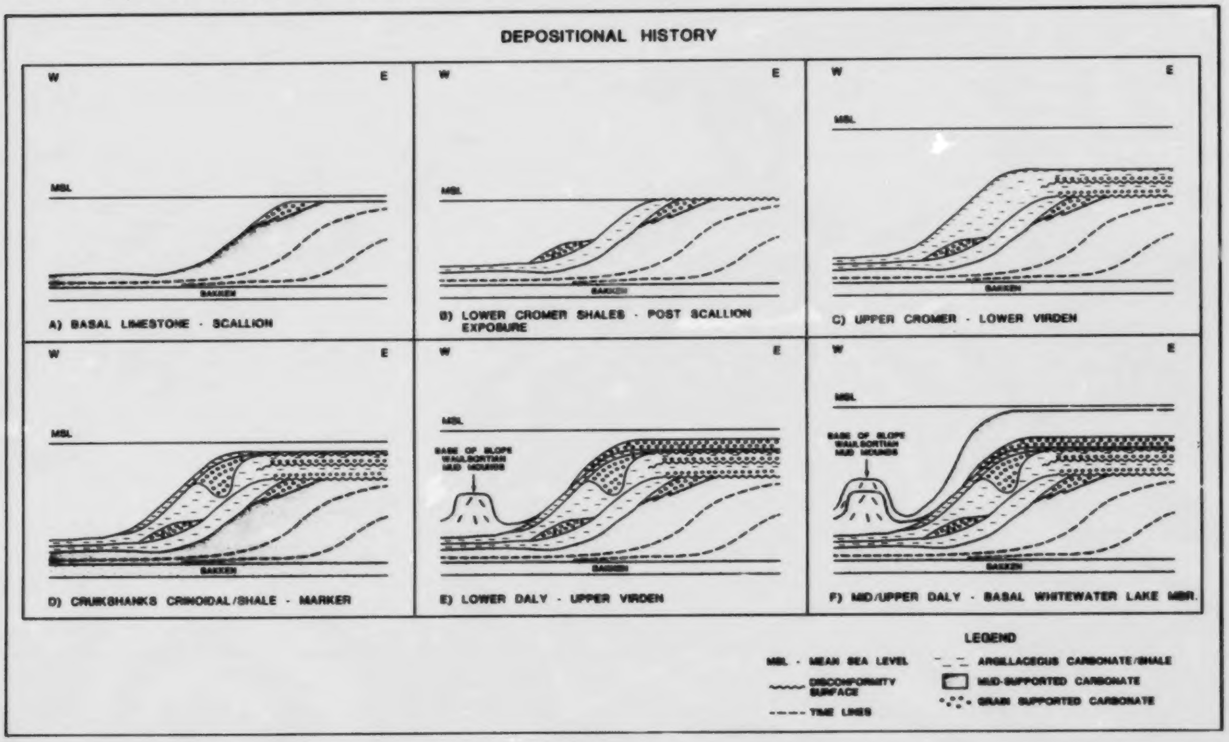
DEPOSITIONAL HISTORY
W
E
MSL
BAKKEN
A) BASAL LIMESTONE - SCALLION
B) LOWER CROMER SHALES - POST SCALLION EXPOSURE
C) UPPER CROMER - LOWER VIRDEN
D) CRUIKSHANKS CRINOIDAL/SHALE - MARKER
E) LOWER DALY - UPPER VIRDEN
F) MID/UPPER DALY - BASAL WHITEWATER LAKE MBR.
BASE OF SLOPE WAULSORTIAN MUD MOUNDS
LEGEND
MSL - MEAN SEA LEVEL
DISCONFORMITY SURFACE
TIME LINES
ARGILLACEOUS CARBONATE/SHALE
MUD-SUPPORTED CARBONATE
GRAIN SUPPORTED CARBONATE

Figure 11: Depositional history and sequence stratigraphic framework of the Lodgepole Formation in the Virden-Daly area.

related to topography on the Paleozoic unconformity surface. The stratigraphic correlations and depositional model presented in this study have several important implications for exploration programs directed at the Lodgepole Formation:

1) The primary control on the location of oil fields is the distribution of oolitic and/or bioclastic grainstones within the middle part of the Lodgepole interval. In the Virden area, the thin but areally extensive grainstone/packstone facies of the Virden Member record episodic sea- level fluctuations across a broad, flat shelf. In the Daly and Kirkella areas, thick accumulations of porous grain-supported lithologies (Cruickshank Crinoidal) are locally developed in relatively thick channels which cut down into slope and base-of-slope deposits of the underlying Cromer Shale. A thinner but more areally extensive grain-supported shoal facies (Lower Daly Member) is correlated with a similar facies in the Upper Virden of the Virden area.

2) Structure and porosity distribution also exert strong control on hydrocarbon distribution. Grainstone lithologies extend beyond the present limits of both the Daly and Virden pools, but are typically tightly cemented between, and along the flanks of, the structural culminations which host the Virden and Daly fields. The accumulation of hydrocarbons apparently prevented occlusion of primary porosity by late calcite/quartz cements.

3) The regional trend of the shelf-slope break (marked by the concentration of shales within the Lodgepole) is oriented north-south in southwestern Manitoba and is approximately parallel to major salt-solution features and basement lineaments such as the Birdtail-Waskada Axis. Regional work by McCabe (1959) suggests that the belt of thick shales within the Lodgepole swings westward into Saskatchewan (Fig. 12). Although no "Virden-type" shelf successions have been identified in cores or logs in Saskatchewan, a number of wells in the Indian Head area (east of Regina) exhibit log/core characteristics similar to wells in the Daly area. Exploration in this area should be focused on highs on the Paleozoic unconformity surface located downdip of major salt collapse features which could provide structural closure for hydrocarbon accumulations.

CONCLUSIONS

1) The Early Mississippian (Kinderhookian) Lodgepole Formation is a complex succession of shallow marine carbonates and shales which host significant reserves of light oil in complex stratigraphic/structural traps along the northeastern rim of the Williston Basin.

2) In the Virden area, several distinct reservoir facies are developed within the Lodgepole. The most prolific reservoir consists of cherty lime mudstones (Scallion Member) which have been pervasively leached at an intra-Mississippian exposure surface. However, large oil reserves are also hosted in laterally extensive oolitic and bioclastic grainstones and packstones of the overlying Virden Member. These beds record episodic sea level fluctuations across a relatively flat, shallow shelf. Farther west in the Daly area, a succession of limy shales containing abundant fenestellid bryozoa and crinoid detritus record slope and base-of-slope sedimentation. Locally, these shales are cut out by wackestone-packstone-filled channels which apparently record a major drop in relative sea level. In eastern Saskatchewan, equivalent strata consist of non-argillaceous lime mudstone which locally form Waulsortian-type mudmounds.

3) The shelf/slope break recognized in the Virden-Daly area of southwestern Manitoba trends approximately north-south. This break is coincident with deep-seated basement lineaments which have localized dissolution of Middle Devonian salts and influenced the geometry of the Lodgepole basin. This shelf/slope break appears to swing westward into Saskatchewan, so undiscovered hydrocarbon accumulations analogous to Virden or Daly may possibly exist along the Lodgepole subcrop edge east of Regina.

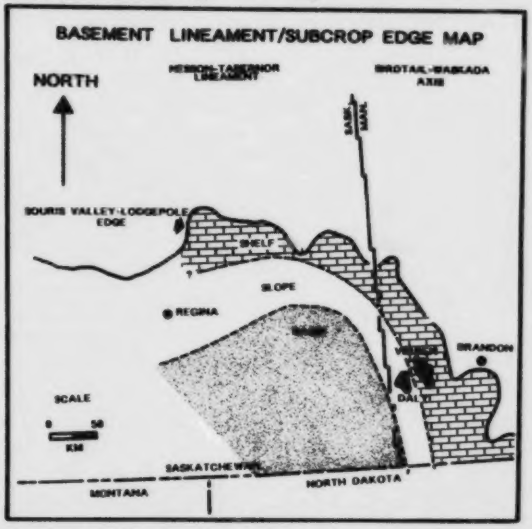

Figure 12: Regional paleogeographic map of the Middle Lodgepole interval along the northeast rim of the Williston Basin. Note the interpreted shelf/slope break trends north-south in Manitoba but swings westward into Saskatchewan. The location of the slope facies in Saskatchewan is marked by the position of shale "thicks" identified within the Lodgepole in this area by McCabe (1959).

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DEPOSITION AND DIAGENESIS OF THE MC-3 MEMBER OF THE MISSION CANYON FORMATION, PIERSON FIELD, MANITOBA

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ABSTRACT

The Mississippian Mission Canyon Formation is a prolific oil reservoir in the Williston Basin. The uppermost member of the formation, the MC-3 unit, is productive in the Pierson field in southwestern Manitoba. In the Pierson area, the MC-3 rocks are characterized by four distinct lithofacies: a) algal laminite/stromatolite facies; b) bioclastic grainstone facies; c) peloidal grainstone facies; and d) vadolite facies. Early diagenesis of the unit consisted mainly of biological micritization and burrowing, marine and freshwater calcite cementation, and dolomitization. Later diagenesis was dominated by dolomitization, anhydritization, compaction and pressure solution, and leaching. Porosity development in the MC-3 member in the Pierson field, like many other Mission Canyon reservoirs elsewhere in the Williston Basin, represents a complex interplay of late-stage fabric-selective dissolution processes coupled with development of intercrystalline porosity and preservation of primary intergranular porosity and diagenetic features.

INTRODUCTION

The Pierson oil field is located in the extreme southwest corner of the province of Manitoba, Canada, about 20 km north of the Canada-United States border, and about 10 km east of the Manitoba-Saskatchewan border (Fig. 1). The 07 43B pool is the largest of the Pierson field pools, and is within the subcrop of the Mississippian Mission Canyon Formation. The producing horizons of the 07 43B pool are found in the MC-3 Member of the formation.

McCabe (1959) included the Pierson area in his regional stratigraphic study of the Mississippian sequence in Manitoba. Later he discussed the general stratigraphic and structural factors responsible for oil accumulation in the MC-3 beds in the Pierson field (McCabe, 1963). More recently, Husain and Halabura (1984) provided a detailed stratigraphic study and assessment of the petroleum geology of the MC-3 Member in this area.

Previous work on the Mission Canyon sequence in adjoining areas of the Williston Basin has given rise to a number of different scenarios regarding porosity development in the unit. Gerhard *et al.* (1978) suggested that porosity and permeability of the stratigraphically equivalent Frobisher-Alida beds in the Glenburn field, North Dakota, are associated with an early diagenetic strandline vadose pisolite as opposed to previously held ideas that the reservoir was a submarine oolite shoal. Elliot (1982), working on the Mission Canyon reservoir in the Haas field, North Dakota, and Bell (1986), examining the Tilston field, Manitoba, concluded that most of the porosity was secondary or secondarily enlarged, and resulted from leaching of calcite during late diagenesis. In their studies on the Little Knife field, North Dakota, Lindsay (1982), Lindsay and Roth (1982), and Lindsay and Kendall (1985) related the development of porosity in the Mission Canyon to early-diagenetic anhydrite replacement of skeletal fragments followed by dolomitization of the muddy matrix, and, finally, late-diagenetic anhydrite leaching. These contrasting scenarios may reflect either regional differences or different interpretations of similar processes due to insufficient data.

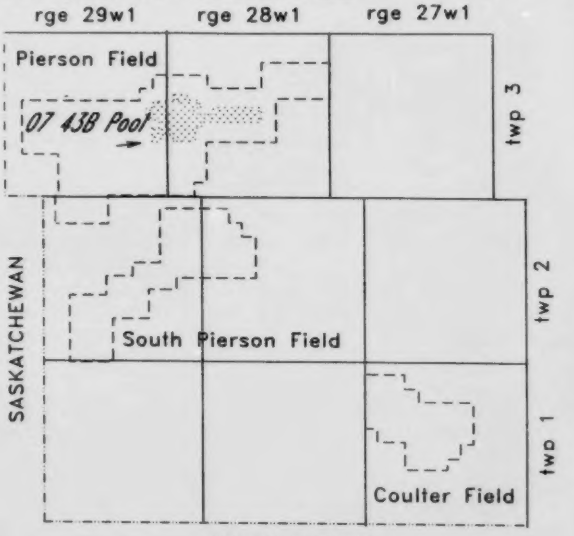


Figure 1: Location map of the Pierson field, southwestern Manitoba.

The purpose of this paper is to summarize the stratigraphic setting and depositional/diagenetic environments of the MC-3 unit and to explain the diagenesis and porosity development of the reservoir interval in the 07 43B pool of the Pierson field in southwestern Manitoba.

METHODS OF STUDY

All available cores from the MC-3a B pool (up to 1989) were examined, logged, and subsampled. A total of 144 subsamples from five cores were analyzed using standard X-ray diffraction, stained thin sections, and scanning electron microscopy-energy dispersive X-ray techniques. Detailed lithologic descriptions and discussion of the analytical techniques and data base are included in Ahmed (1985).

STRATIGRAPHIC SETTING AND DEPOSITIONAL ENVIRONMENT

Southwestern Manitoba is located on the northeastern edge of the Williston Basin. The basin has been tectonically active from Ordovician to Cretaceous time and was the principal feature governing the pattern of sedimentation in the region during this interval (McCabe, 1971).

Two gross lithologic sequences are present in southwestern Manitoba: a) the lower Paleozoic sequences composed almost entirely of carbonate rocks; and b) the upper Mesozoic and Cenozoic sequences consisting mainly of shale and sandstone. These two sequences are separated by a pronounced angular unconformity, with Mississippian and Upper Devonian carbonates buried beneath overlapping Mesozoic strata. Consequently, an irregular series of northwest-trending subcrop belts is developed beneath the Jurassic sediments. Throughout much of southwestern Manitoba, the Paleozoic rocks immediately underlying this unconformity have been extensively modified by post-Mississippian dolomitization and anhydritization creating a regionally extensive "altered" zone which acts as a caprock in many Mississippian oil accumulations.

The Mississippian System in southwestern Manitoba consists of four formations: Bakken, Lodgepole, Mission Canyon, and Charles Formations. The Mission Canyon Formation conformably overlies the Lodgepole Formation and has been subdivided into three members: MC-1, MC-2, and MC-3 Members. The limestone, dolomite, and evaporites of the uppermost MC-3 Member have informally been further subdivided into lower (MC-3a) and upper (MC-3b) units.

The MC-3 beds in southwestern Manitoba generally fall within the restricted shallow marine platform facies of Wilson (1975). This parallel-to-shore facies belt contains sediments spanning a considerable spectrum of carbonate tidal flat and shallow marine environments, including shoals, tidal channels, cut-off lagoons, and coastal ponds with restricted circulation.

MC-3A LITHOFACIES IN THE PIERSON AREA

This study has recognized four basic groups of lithofacies in MC-3a rocks in the Pierson area. These are: a) stromatolite, cryptalgal laminite, and pseudostromata facies; b) bioclastic grainstone facies; c) peloidal grainstone facies; and d) vadolite facies. Their distribution and thickness are extremely variable. Consequently, detailed lithostratigraphic correlation, even over small distances, is difficult. The thick algal facies, which has been found in all cores examined, constitutes the main rock type of the area. Other facies show thin and sporadic development. Figure 2 provides a generalized distribution of lithofacies in a 10 km^2 area of the field.

The variability of distribution and thickness of the lithofacies indicates an ephemeral depositional environment which Lobue (1982), describing the Silurian Interlake Formation in North Dakota, proposed was related to: a) active tidal channels; and b) a shallow depositional slope. These two factors allow even minor fluctuations in relative sea level to cause extensive physiographic changes giving rise to a dynamic depositional system.

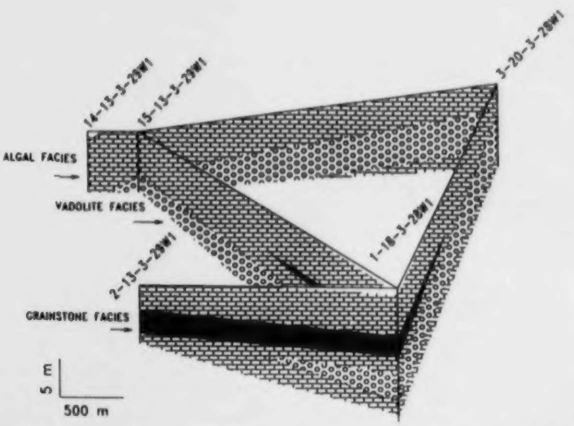


Figure 2: Panel diagram showing lithofacies distribution in the 07 43B pool area of the Pierson field.

a) Stromatolite/Cryptalgal Laminite/Pseudostromata Facies

This facies comprises the main rock type of the MC-3 Member in the Pierson area. All rocks of this facies are algal in origin and have been subdivided into three subfacies on the basis of the particular growth form of the algae: i) stromatolite subfacies; ii) cryptalgal laminite subfacies; and iii) pseudostromata subfacies.

The stromatolite subfacies is dominated by LLH type (laterally linked hemispheroids) and SH type (stacked hemispheroids) structures, and is characterized by both irregular and laminoid fenestrae between the stromatolitic laminae. It is an algal mat lofrite as both fenestrae and sheet cracks are developed. Fossils are scarce, represented by ostracode, dasyclad, foraminifera, and brachiopod bioclastic debris. Commonly, pockets of peloids are entrapped in the mat.

The cryptalgal laminite subfacies is composed mainly of micrite and algal peloids bounded by blue-green algae. Laminae exhibit both planar and irregular anastomosing fabrics. In addition to minor bioclastic debris, the sediment contains terrigeneous argillaceous components.

The pseudostromata subfacies consists of irregular, dense patches of granuloid algal micrite with well defined fenestrae, sheet cracks, vertical desiccation cracks and vertical burrows. In all of the cores examined in the Pierson area, this subfacies is associated with the stromatolite subfacies, suggesting that it may represent stromatolites homogenized by bioturbation.

Interpreted depositional environments of these various subfacies range from protected intertidal flats and pools to exposed intertidal mudflats.

b) Bioclastic Grainstone Facies

The bioclastic grainstone facies is composed of mainly Cyanophycean algal grains (e.g., *Girvanella*), foraminifera, dasyclads, echinodermal fragments, and bryozoa (Fig. 3). Brachiopods, red algae, and rugose coral fragments are also present in minor amounts. The facies has limited spatial distribution and was observed in only one core. It probably represents deposition in a restricted marine shoal area.

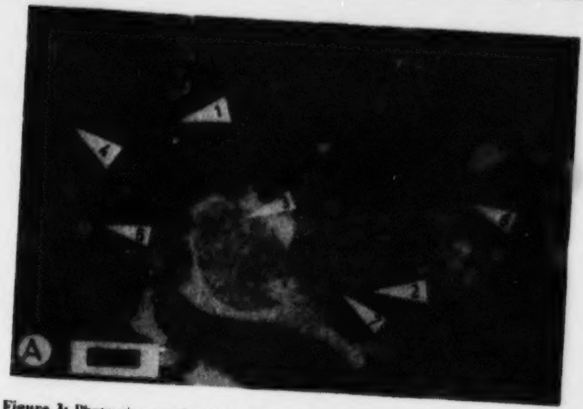


Figure 3: Photomicrograph of bioclastic grainstone facies showing (1) algal grains, (2) foraminifera, (3) echinoderm fragments, (4) meniscus cement, (5) modified interparticle porosity, (6) vuggy porosity, and (7) intraparticle porosity. Scale bar is 0.25 mm; crossed nicols; sample from 986.0 m depth in 1-18-3-23W1.

c) Peloidal Grainstone Facies

The peloids making up this rock consist of fecal pellets, rounded and sorted aggregates of carbonate mud (pseudopelletoids), and micritized, rounded and abraded bioclasts. The bioclasts include echinodermal fragments, dasyclads, and foraminifera. This rock type was probably deposited in restricted marine shoal areas.

d) Vadolite Facies

The term vadoids refers to coated grains which originated in a vadose environment due to non-biological processes (Peryt, 1983). Vadoids are synonymous with caliche pisoids, vadose pisoids/ooids, and cave pearls. Environments of formation of vadoids and the vadolite facies include freshwater, marine, and hypersaline settings.

The vadolite facies in the MC-3a sediments of the Pierson area consists mainly of coated grains and crusts with variable amounts of micritic matrix. The sediment is well packed and grain supported with rare occurrences of vadolites in wackestones. The grains have irregular spherical shapes and range in diameter from less than 2 mm to greater than 10 mm. Both sorting and grading are common; inverse graded beds up to 10 cm thick are present. The vadolites themselves are usually composed of alternating laminae of micrite and radiaxial spar. Nucleii are generally nondescript micritic pellets or fragments of broken vadoids; more rarely, they are bioclasts of ostracodes, foraminifera, gastropods, or dasyclads. The vadoids are commonly fractured, presumably due to desiccation, with many showing multiple stages of fracturing, coating, and refracturing (Fig. 4). Many of the grains of the vadolite facies are flattened and some show a distinctive polygonal "fitted" pattern. Some vadolites show a downward thickening of the cortex and many share outer laminae. In many places, these outer laminae of vadolites continue into the surrounding sediment and locally merge into crusts and accretionary cements (Fig. 5). Where appropriate contacts are preserved in core, the vadolite facies grades into the algal facies.

The vadolite facies is interpreted to have formed in an overall marine vadose environment, and is probably the result of exposure and very early diagenesis of the ubiquitous lofritic algal mat. The rather sporadic distribution of the facies in the Pierson area suggests that vadose conditions existed only locally, depending on geomorphology and amplitude of sea level fluctuation.

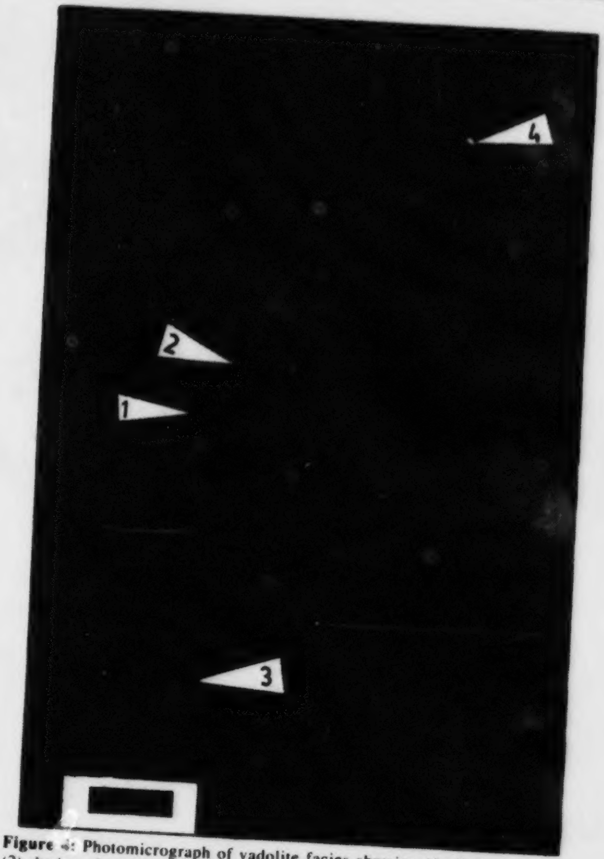


Figure 4: Photomicrograph of vadolite facies showing (1) flattened grains, (2) sharing of outer accretionary laminae, (3) vadoids having broken vadoids as nucleii, (4) multistage grains, polygonal filling, and vertical asymmetry of grains. Scale bar is 1 mm; sample from 980 m depth in 3-20-3-28W1.

Figure 5: Photograph of vadolite facies showing (1) vadose crust, and (2) inverse grading of the vadolites. Scale bar is 1 cm; sample is a polished slab perpendicular to stratification from 991.8 m depth in 15-13-3-28W1.

DIAGENESIS

BIOLOGICAL DIAGENESIS

Boring endolithic algae and fungi, and burrowing organisms have been instrumental in early biological diagenesis in the MC-3 unit, resulting in micritization and sparmicritization, and in the formation of burrows and peloids. Micrite coatings around grains are present in all the facies types. However, completely micritized grains which possess no traces of skeletal microstructure are more common and are particularly abundant in the peloidal grainstone facies. Sparmicritization (conversion of sparry calcite cement to micrite by degrading recrystallization) is observed in the algal and vadolite facies, and probably took place in the vadose zone as early calcite cement which filled fenestrae and desiccation voids was affected by boring algae. Discrete burrows have been observed in only the pseudostromata subfacies. These burrows show dominantly vertical orientations.

CALCITE CEMENTATION

Calcite cement of early diagenetic origin is one of the major porosity reducers as well as a principal contributor to lithification of the MC-3 beds in the Pierson area. Both marine and freshwater cements are present.

Radiaxial calcite cements are most common in the bioclastic grainstone and vadolite facies. These inclusion-rich cements are characterized by bladed morphology, alignment of C-axis perpendicular to the pore-boundary, and undulose extinction (Fig. 6). They are probably a neomorphic alteration product of a fibrous aragonite precursor which was subjected to aggrading recrystallization in hypersaline waters. This type of cement forms in either a marine phreatic or marine vadose environment both of which are represented in the MC-3 Member.

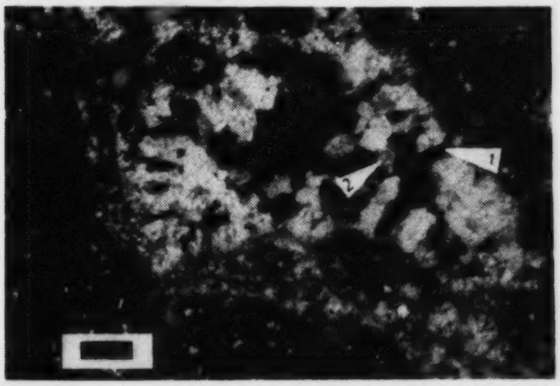


Figure 6: Photomicrograph showing bryozoa bound by isopachous radiaxial calcite cement (1). This cement is overlain by equant calcite cement (2). Scale bar is 0.25 mm; crossed nicols; sample from 984.5 m depth in 1-18-3-23W1.

Marine phreatic cement is found only in the bioclastic grainstone facies and forms only a minor percentage of the overall cement component of that facies. A phreatic origin is inferred by the isopachous distribution of the cement around porewalls. It mainly occurs in intraparticle pores of bioclasts.

Marine vadose cement is present only in the vadolite facies. The vadolites were formed by accretionary growth of alternating laminae of micritic calcite and fibrous aragonite which later inverted to radiaxial calcite. Radiaxial calcite is also found in the vadolite facies forming microstalactitic cements.

In addition to the isopachous and microstalactitic marine cements, various types of freshwater cements are present in the MC-3 beds. One of the most common is a meniscus cement. It is found only in the bioclastic grainstone facies and exhibits a patchy, uneven habit (as opposed to the even cementation characteristic of the marine phreatic zone). The freshwater meniscus cements are preferentially developed in small pores (Fig. 7).

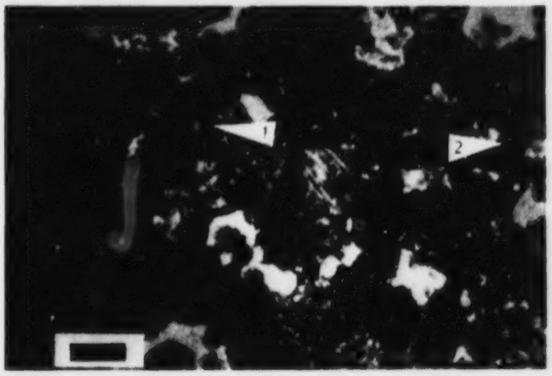


Figure 7: Photomicrograph showing meniscus cement preferentially developed in small interparticle pores (1) as opposed to the larger, more easily drained pores (2). Scale bar is 0.5 mm; crossed nicols; sample from 986.2 m depth in 1-18-3-23W1.

Finally, both calcite syntaxial overgrowths and equant cements are found in the rocks. Syntaxial overgrowths, typically developed around echinoderm fragments, are most commonly present in the bioclastic grainstone facies. This cement fills large areas of the pore space. Equant calcite cement is developed mainly in the fenestrae and desiccation cracks of the stromatolite and pseudostromata subfacies, and rarely occurs in the bioclastic grainstone facies. This pore-filling cement shows coarsening toward the centre of the pore, suggesting formation in a phreatic environment.

All calcite cements observed in the Pierson area appear to be products of early diagenesis before burial. The most definitive sequence of calcite cementation events is observed in the bioclastic grainstone facies. Here vadose meniscus cements predate syntaxial overgrowths which, in turn, are enclosed by equant calcite cements. In contrast, marine phreatic cement, when present, appears to be partially leached, then overlain by calcite equant cement.

NEOMORPHISM

With the advent of a freshwater phreatic environment, inversion of aragonite which survived leaching and recrystallization of high-magnesian calcite gave rise to calcite as a neomorphic product. As the freshwater phreatic environment persisted, some micrite suffered aggrading neomorphism. This aggrading neomorphism is sparsely developed, occurring mainly in the algal facies where micritic algal filaments are converted into microspar and pseudospar. In contrast, aggrading neomorphism of early marine fibrous aragonite cements in the vadolite facies took place under saline and hypersaline conditions, and resulted in the formation of radiaxial calcite cement. The relative timing of this is not evident from the fabric of the rock.

DOLOMITIZATION

The altered zone is a regionally extensive lithostratigraphic unit immediately underlying the Mississippian-Jurassic unconformity. Numerous studies (e.g., McCabe, 1959, 1963; Kendall, 1975; Young and Greggs, 1975; McCulloch-Smith, 1984; Rodgers, 1986) have noted that the thickness of the altered zone is inversely related to the thickness of the overlying Jurassic red bed sequence: where the Jurassic sediments are thin, a thick altered zone is present in the Mississippian sediments.

The altered zone in the Pierson area is of variable thickness, but averages about 3 m. It is readily identified on wireline logs by a low sonic interval transit time and a high resistivity kick. It consists of pervasively dolomitized sediment with abundant anhydrite stringers and nodules. Microscopic examination shows that dolomite occurs mainly as fine- to medium-crystalline, anhedral to subhedral rhombs which characteristically contain abundant inclusions of black, insoluble residue and some of which are discernibly zoned; optical extinction is diffuse.

Dolomite also occurs in relatively minor amounts in the MC-3 unit in the Pierson area. It has three basic forms: a) microcrystalline material dispersed throughout the carbonate mud and in small patches within the micritized grains and carbonate cements; b) fine- to medium-crystalline, clear to white, sparry limpid rhombs; and c) incipient rhombs with very faint extinction and a granular appearance. Mattes and Mountjoy (1980) reported this last form which they termed mosaic dolomite.

The variety of dolomite types, the pervasive nature of the mineral in the altered zone compared with its sporadic and diffuse occurrence in underlying sediments, and its association with grains, mud, and cement, together suggest that dolomite is polygenetic in origin. The relatively small amount of microcrystalline dolomite is likely the result of penecontemporaneous dolomitization of calcite or aragonite precursors. Much of this early replacement product has undergone subsequent neomorphism to give rise to the coarser crystalline dolomite. The principal phase of dolomitization of the MC-3 carbonates probably occurred during formation of the altered zone, most likely related to downward percolation of hypersaline brines as overlying Jurassic red beds were being deposited (Fig. 8).

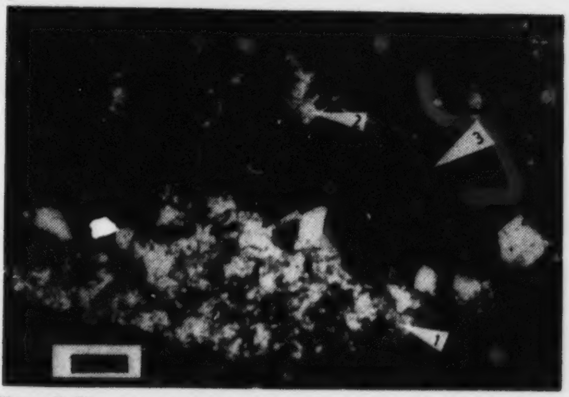


Figure 8: Photomicrograph showing late-stage geopetal (1) and pore-lining (2) dolomite in a vadomoldic pore (3) Scale bar is 50 microns; crossed nicols; sample from 991.8 m depth in 1-18-3-23W1.

DEDOLOMITIZATION

In the calcitic dolostone immediately below the altered zone, calcite occurs as dispersed crystals which enclose corroded dolomite rhombs. Many of these calcite crystals appear to be pseudomorphs after dolomite rhombs, indicating an episode of calcite replacement of dolomite (dedolomitization) by high Ca/Mg pore fluids. Dedolomitized sediments are restricted to this geological setting.

ANHYDRITIZATION AND GYPSIFICATION

Anhydrite is a common diagenetic mineral throughout the MC-3 sequence in the Pierson area. It is most abundant in the altered zone where it occurs mainly as bluish grey to white bands and stringers up to 15 cm thick. Although relatively free of carbonate, irregular patches, blebs, and stringers of host dolomite occur in this anhydrite. Anhydrite is also present in sediments underlying the altered zone as: i) fracture-filling material; ii) moldic pore-filling cement; and iii) xenotopic to poikilotopic replacement of matrix and allochem material.

Gypsum is associated with the anhydrite and is generally distinguishable from it only by X-ray diffraction. On the basis of XRD examination, gypsum makes up 15 percent of calcium sulphate in the MC-3a sediment. Gypsum also occurs in the vadolite facies as coarse, tabular to fibrous crystals commonly exhibiting swallow-tail twinning.

FORMATION OF IRON MINERALS

Hematite, limonite, and pyrite are present in small amounts disseminated throughout the MC-3 sediment. Black to red hematite and amorphous nodules of brassy yellow pyrite also occur as cumulates along stylolites. Pyrite is occasionally found filling fractures and replacing bioclasts. Similarly coloured material identified as marcasite, identified by X-ray diffraction analysis, occurs in fractures and lining the walls of vugs. The petrographic relationships and time of formation of these various secondary iron precipitates in the Mission Canyon Formation are obscure. Most of the hematite and pyrite appear to have formed relatively early in the diagenetic history, before substantial burial, pressure solution and compaction took place. However, the fracture-filling and fracture-lining habit of some pyrite and marcasite indicates their late-stage genesis.

FRACTURING

All cores examined show fractures which are mainly vertical, having lengths up to one metre. All of the fractures in the altered zone are infilled with anhydrite, whereas the section below the altered zone exhibits open fractures, fractures with oil stain and fractures partially infilled with pyrite. Although fracturing is obviously a post-lithification event, its relative timing in the Pierson area is not evident from available petrographic relationships.

PRESSURE SOLUTION

Stylolites have developed within all the facies as well as along facies boundaries in the section below the altered zone. No stylolites have been observed in the altered zone. The presence of dolomite, pyrite and hematite as cumulates, and the occurrence of dead oil along the stylolites indicate that stylolite formation took place relatively late in the diagenetic history of the unit.

POROSITY TYPES

Porosity of the MC-3a beds in the Pierson area ranges from less than 5 percent up to 18 percent. The main types of porosity (Fig. 9) are: i) interparticle voids, dominant in the vadolite, peloidal grainstone, and bioclastic grainstone facies; and ii) fenestral porosity which occurs in the algal facies. Intercrystalline porosity is also present in the sediments immediately underlying the altered zone. Fabric-selective biomoldic and vadomoldic porosity occur in the grainstone and vadolite facies (Fig. 10). Finally, leached intraparticle and fracture porosity occur in minor amounts in all of the lithofacies.

Analysis of quantitative X-ray diffraction and thin-section data collected from the 144 subsamples of the MC-3a sediments show that no statistically significant relationship exists between the amount of porosity and the degree of dolomitization or anhydritization of the rocks. Although, in general, the vadolite sediments show slightly higher porosity values, statistically significant differences between the various lithofacies were not found.

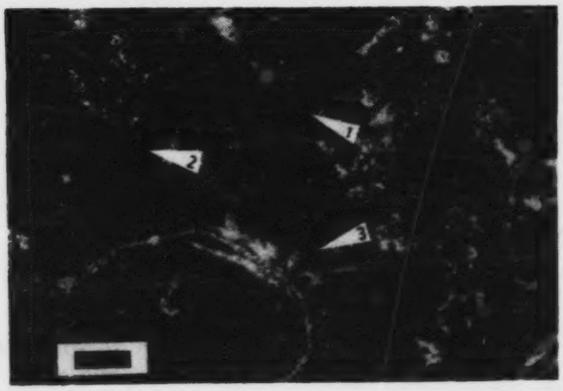


Figure 9: Photomicrograph of the vadolite facies showing (1) modified interparticle porosity, (2) vuggy porosity, and (3) channel porosity. Total porosity of the sample is 13.5 percent. Scale bar is 0.5 mm; crossed nicols; sample from 990.9 m depth in 15-13-3-28W1.

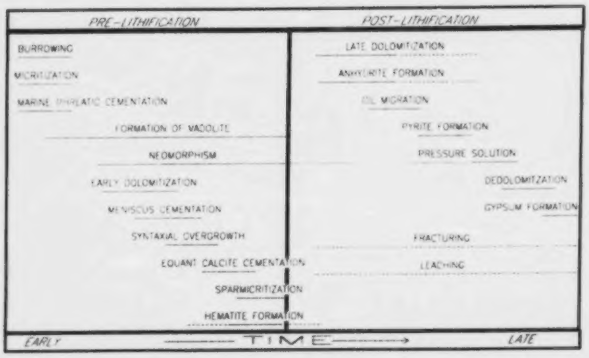
Figure 10: Photomicrograph showing vadomoldic and biomoldic porosity. Scale bar is 30 microns; crossed nicols; sample from 982.9 m depth in 3-20-3-28W1.

SUMMARY

In the study area, the MC-3a beds represent a carbonate tidal flat-shoreline complex. Four distinct facies can be recognized in cores: algal, bioclastic grainstone, peloidal grainstone, and vadolite facies. The vadolite facies most likely formed during subaerial exposure and very early diagenesis of the algal mat-stromatolitic sediments.

The large number of diagenetic environments, the variability of products in each, and the repetitive nature of preserved strata make chronological interpretation of diagenetic events affecting the Mississippian sediments in this area complex. Sufficient evidence exists for deciphering major events from which a workable diagenetic sequence has been established (Table 1).

Table 1: Relative chronology of the most important diagenetic events affecting the sediments of the MC-3 unit in the Pierson area.



Diagenetic processes took place in two stages: early (pre-lithification) and late (post-lithification).

Early diagenesis is characterized mainly by biological activity, compaction, formation of vadolite, and calcite cementation. Syndepositional burrowing in shallow water algal carbonates of the MC-3a unit resulted in the formation of pseudostromata. Periodic exposure and development of a freshwater phreatic environment resulted in dissolution of much of the metastable aragonite and Mg-calcite or recrystallization of these carbonates to calcite in the algal facies. Equant calcite cements formed below the watertable whereas sparmicritization took place in the freshwater vadose environment. Subaerial exposure of these shallow-water algal sediments was also responsible for formation of the vadolite facies.

In slightly deeper water sediment, the bioclastic and peloidal grainstone facies were first affected by micritization of bioclasts and the formation of marine cement. The succeeding freshwater phreatic environment resulted in the precipitation of syntaxial overgrowth cement followed by equant calcite cement. The change from a marine to freshwater setting also brought about dissolution and/or neomorphism of Mg-calcite and aragonite into calcite.

Late diagenesis includes dolomitization, anhydritization, formation of authigenic iron minerals, compaction, pressure solution, fracturing, and leaching. The relative chronology of these late-stage diagenetic events generally shows considerable overlap. Most dolomitization and anhydritization were apparently due to downward percolating hypersaline brines associated with deposition of the overlying Amaranth redbeds.

The development of porosity is not truly facies selective. Late diagenetic processes controlled both the preservation of primary porosity and the development of leached porosity and minor intercrystalline and fracture porosity.

ACKNOWLEDGEMENTS

This petrographic study of the MC-3 unit in the Pierson area was supported by funds from the Canadian Commonwealth Scholarship and Fellowship program and the University of Manitoba. Manitoba Energy and Mines kindly provided access to the core storage facility at the University of Manitoba. We are grateful to Drs. G. Clark and A. Baracos of the University of Manitoba for their helpful suggestions and critical comments.

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MISSISSIPPIAN OIL PRODUCTION IN PART OF BURKE, MOUNTRAIL AND WARD COUNTIES, NORTH DAKOTA

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ABSTRACT

Important oil producing units in southern Burke, northern Mountrail and northeastern Ward counties generally are below the State A marker in the Coteau, Bluell, and Sherwood beds of the Mississippian Madison Group. Eleven oil fields in this area have produced from one or more of these units. To the north of this area, oil producing units are stratigraphically higher, primarily the Mississippian Rival subinterval of the Frobisher-Alida interval and the Midale subinterval of the Ratcliffe interval of the Madison Group.

Oil production generally occurs below northeast to southwest trending elongated thickenings, or "noses", of the Ratcliffe interval. These noses are clearly seen on isopachous maps of the Ratcliffe. Another useful indicator of these noses is the edge of the F salt. The salt edge generally trends northwesterly and diagonally across this area, and outlines the noses.

INTRODUCTION

Oil production on the eastern side of the Williston Basin in North Dakota (Fig. 1) is primarily from Mississippian formations belonging to the Madison Group (Fig. 2a). Burke, Mountrail and Ward counties are located within the eastern portion of the Basin (Fig. 1) and have produced millions of barrels of oil from Madison reservoirs.

Early production in north central Burke county during the late 1950's was from various subintervals of the Charles Formation, particularly the Midale and Rival (Nesson). Recent drilling to the Mission Canyon Formation has tapped reservoirs at greater depths, and the volume of oil produced from Mission Canyon subintervals has surpassed that from the shallower subintervals of the Charles. This deeper drilling in north central North Dakota resulted in discovery of oil in the Coteau subinterval and hence the naming of Coteau Field. The adjacent Flaxton Field also produces from the Coteau and the reservoir characteristics have been well documented (Voldseth, 1986; Voldseth, 1987).

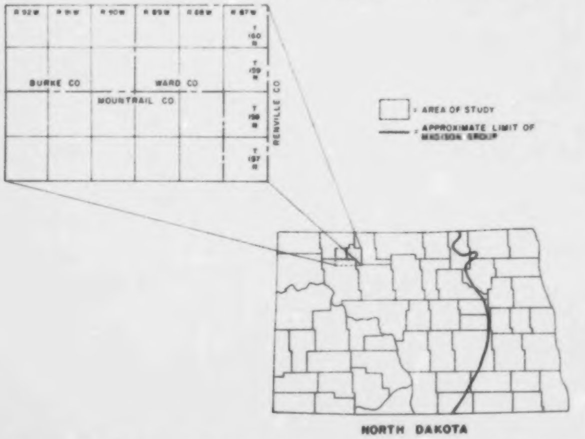

Figure 1. Location map of the area studied for this report. The approximate limit of deposition of the Madison Group is shown.

Production from the Midale and Rival subintervals in north central Burke County was shown by Anderson *et al.* (1960) to be localized below elongate northeast to southwest paleostructures of the Ratcliffe interval. Anderson *et al.* (1960) showed that the Ratcliffe paleostructures could be quickly mapped by mapping the edge of the F salt. The F salt edge in north central Burke County continues to the southeast into southern Burke, northern Mountrail, and northeast Ward counties, and coincides with the

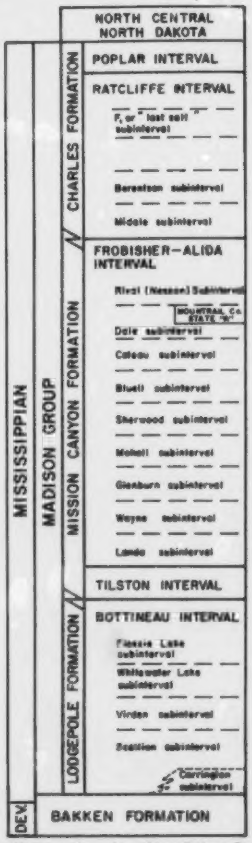

Figure 2a. Nomenclature and stratigraphic relations for the units discussed.

location of production in subintervals of the Mission Canyon Formation, but currently not the Midale or Rival. Because the majority of wells in this area produce oil from the Coteau subinterval of the Mission Canyon, the Coteau is the focus of this paper.

LOCATION AND STRUCTURAL SETTING

The Williston Basin is an oval structural depression about 345,000 km^2 (133,000 mi^2) beneath the northern Great Plains of the United States and Canada, in parts of Manitoba, Saskatchewan, Montana, North Dakota and South Dakota (Gerhard *et al.*, 1990). Burke, Mountrail, and Ward counties on the northeastern flank of the Basin in North Dakota (Fig. 1) have produced millions of barrels of oil from over twelve different subintervals of the Madison Group. Subtle anticlinal structures plunging west-southwest along the eastern flank of the Basin localize Mississippian oil reservoirs (Fig. 3) including those in the oil fields reported here (Fig. 3). Eleven fields have produced oil from five different subintervals, but only three of these, the Greenbush, Southwest Aurelia, and Niobe are currently producing, and four have been abandoned.

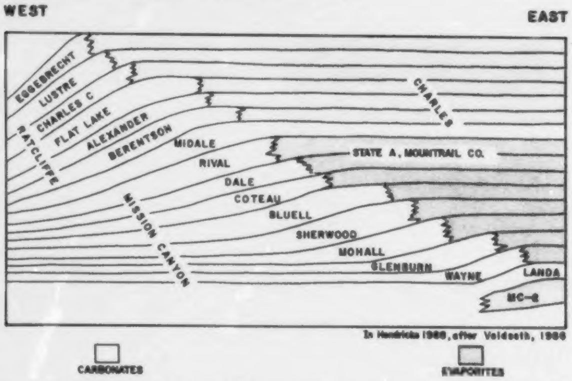

Figure 2b. Generalized schematic diagram of stratigraphic relations as modified from Hendricks 1988 and Voldseth 1986.

STRATIGRAPHY

The Madison Group in North Dakota comprises the Lodgepole, Mission Canyon, and Charles Formations (Fig. 2a). Intercalated beds of basin, shallow shelf, and strand carbonate and evaporite deposits characterize the Mission Canyon and Charles Formations (Fig. 2b). These beds represent pulses of minor sea level rise within a larger regressional event.

Laterally continuous, intercalated units of shallowing upward sedimentary sequences within the Mission Canyon Formation were subdivided into mappable beds (Harris *et al.*, 1966) and intervals (Voldseth, 1986; Figs. 2a and 2b). The nomenclature used in Figure 2 is modified from their work to conform to that used by the North Dakota Geological Survey (Bluemle, *et al.*, 1986). The State "A" log pick corresponds to that used by Anderson *et al.* (1960) in the adjacent townships in north central Burke County. A typical log signature for these units is shown in Figure 4.

COTEAU SUBINTERVAL

The Coteau subinterval was recognized by Voldseth (1986) as a distinct unit that previously had been combined with the upper Bluell subinterval. Voldseth (1986, 1987) subdivided the upper Bluell into the Dale and the Coteau subintervals (Fig. 2). Each of these units consists of an upward shallowing limestone to evaporite sequence.

A structure map of the top of the Coteau (Fig. 3) shows subtle structural flexures, or noses. The axis of these structures trends generally northeast.

Thickness changes in the Coteau subinterval (Fig. 5) are seen to be elongated northeasterly with thick Coteau deposits separated by thins. This pseudoridge and swale system probably reflects the influence of pre-Coteau paleotopography on Coteau depositional environments.

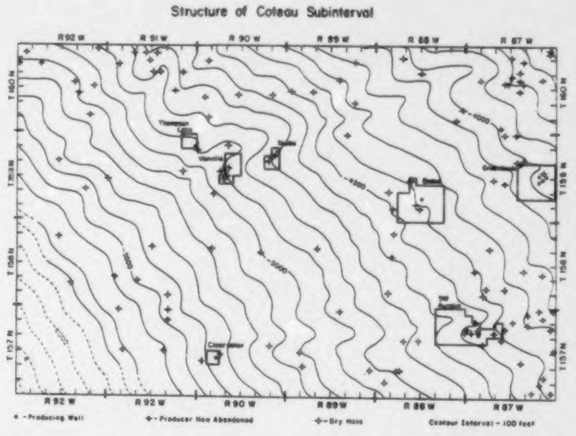

Figure 3. Present structure on the Coteau subinterval with all wells penetrating the Coteau, and their current status. Only oil fields with significant or recent production are shown.

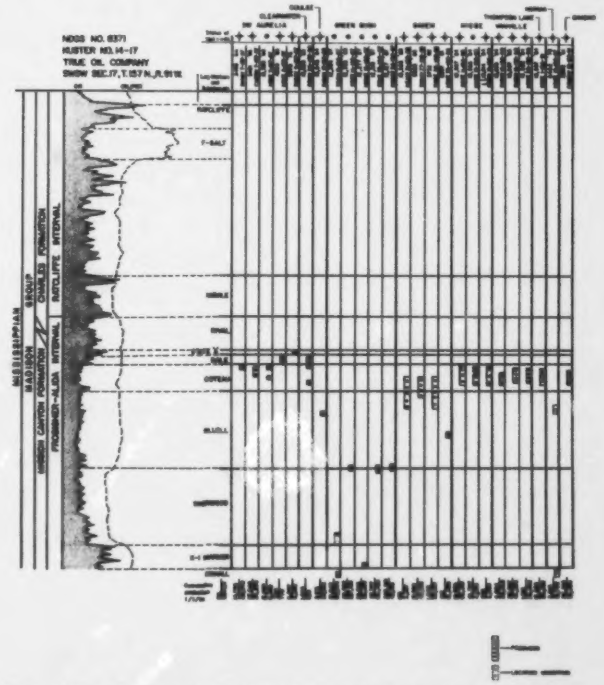

Figure 4. Typical gamma ray log signal in the study area for the Mississippian subintervals. The log picks for the subintervals and log markers are shown. All wells that have produced oil in this area are shown with the producing horizon designated by the perforated zone.

PRODUCTION SUMMARY

Multiple pay zones are characteristic of Mississippian hydrocarbon production in the Williston Basin. This is typified in the study area where production is from five zones. Multiple exploration targets are provided by the many Mississippian subintervals with similar lithologies and textures in vertical sedimentary sequence. Of the five producing zones in this study (Fig. 4), the Sherwood has yielded the most oil and the Dale beds the least. The five producing zones in decreasing order of production are the Sherwood, Coteau, Sherwood/K-1/Mohall, Bluell, and Dale subintervals.

Greenbush field has produced the greatest volume of oil, 410,807 barrels, the larger portion, 262,951 barrels, from the well P.G. & E. Resources Companies, Johnson #4-24, NWNW 24-159-87, NDGS #12,127 in the upper Sherwood subinterval. Greenbush production is 41 percent of the 640,804 barrels produced in the study area.

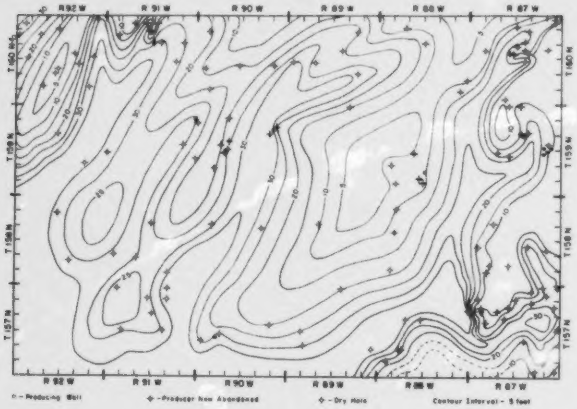

Figure 5. Isopachous map of the Coteau subinterval with all wells penetrating the Coteau and their current status indicated.

The lowest producing field is the Clearwater, a single well field which yielded 1644 barrels of oil. However, like many single well fields in this area, the region surrounding Clearwater has not been thoroughly tested.

The largest number of wells in a field is five in Greenbush and Southwest Aurelia, whereas five of the fields have only one well. The most wells completed in a stratigraphic unit is twelve in the Coteau subinterval. Southwest Aurelia and Niobe (Fig. 6a) fields are the two currently producing from Coteau beds although previously Vanville (Fig. 6b), Thompson Lake, and Dimond fields had Coteau production.

The second largest number of completions in a stratigraphic unit is three and that occurs in each of the Coteau/Bluell, Bluell, and Sherwood subintervals.

The best production has been in the thoroughly explored eastern portion of the study area and from the deeper stratigraphic units. Much of the southwestern corner is unexplored.

COTEAU PRODUCTION

Oil reservoirs in the Coteau subinterval are localized on the flank of structural noses represented by thinning of Coteau beds (Figs. 3 and 5). Structural cross sections of two fields with Coteau production further confirm the flank location of the reservoirs (Fig. 6). It can be seen from the logs on these sections that production is from rocks with limestone matrix porosity of less than 10 percent. Preliminary examination of cores indicates that porosity in producing beds is primarily interparticulate although

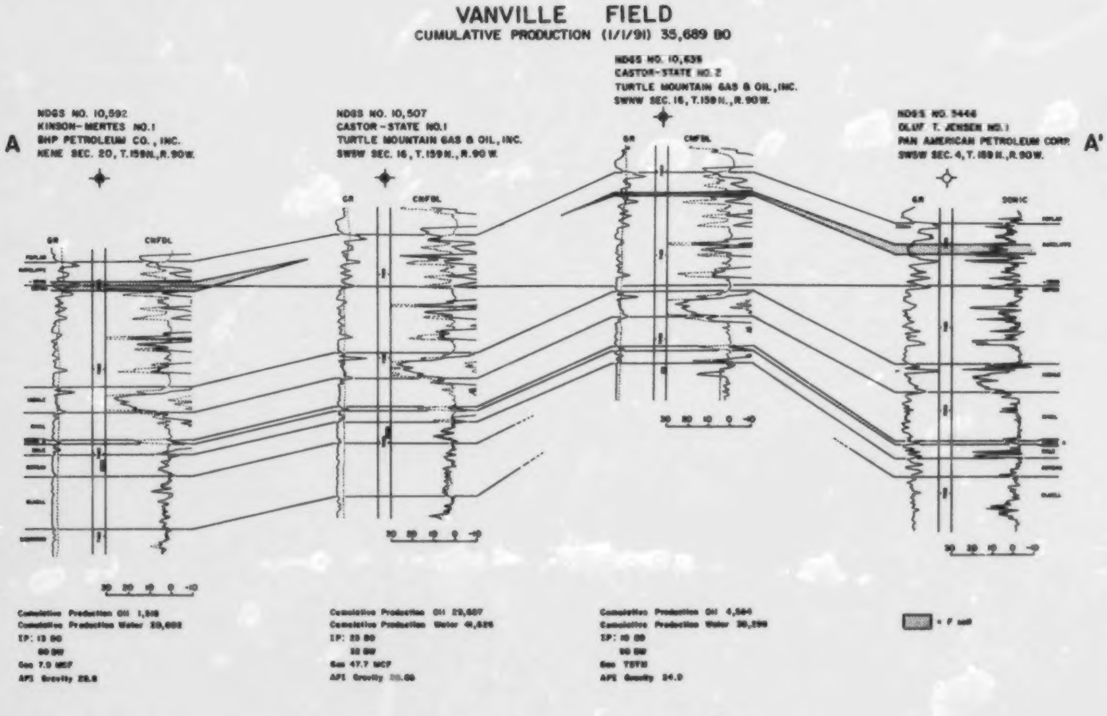
VANVILLE FIELD
CUMULATIVE PRODUCTION (1/1/91) 35,689 BO
A
NDGS NO. 10,592
KINSON-MERTES NO.1
BHP PETROLEUM CO., INC.
NENE SEC. 20, T.159N.,R.90W.
NDGS NO. 10,507
CASTOR-STATE NO.1
TURTLE MOUNTAIN GAS & OIL, INC.
SWSW SEC. 16, T.159N.,R.90W.
NDGS NO. 10,635
CASTOR-STATE NO. 2
TURTLE MOUNTAIN GAS & OIL, INC.
SWNW SEC. 16, T.159N.,R.90W.
NDGS NO. 5444
OLUF T. JENSEN NO.1
PAN AMERICAN PETROLEUM CORP.
SWSW SEC. 4, T.159N.,R.90W.
A'
GR
CNFDL
SONIC
30 20 10 0 -10
Cumulative Production Oil 1,318
Cumulative Production Water 20,002
IP: 13 BO
80 BW
Gas 7.0 MCF
API Gravity 28.8
Cumulative Production Oil 29,807
Cumulative Production Water 4,829
IP: 20 BO
30 BW
Gas 47.7 MCF
API Gravity 30.00
Cumulative Production Oil 4,564
Cumulative Production Water 30,290
IP: 10 BO
90 BW
Gas TSTM
API Gravity 24.0
▨ = F unit

Figure 6a. Structural cross section for Vanville Field (A-A').

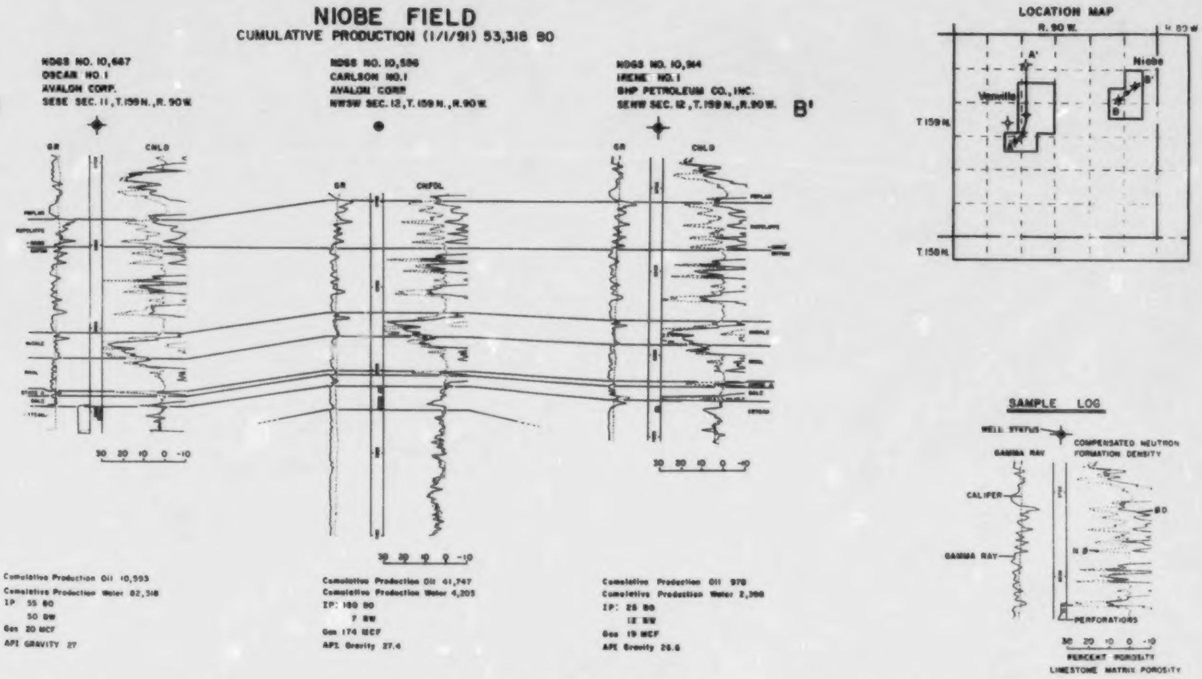

Figure 6b. Structural cross section for Niobe Field (B-B'). Subintervals are correlated and producing unit is seen by the perforated zone to be the Coteau subinterval. Scales for limestone matrix porosity are indicated for the density logs. Production information is current through 1 January 1991.

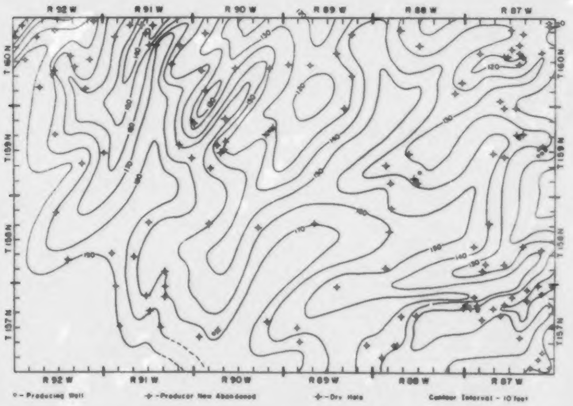

Figure 7. Isopachous map of the Ratcliffe interval showing the elongated thickenings and thinnings of the unit. Note that producing wells are coincident with Ratcliffe paleotopographic highs.

uncemented vertical fractures also may be important. This type of porosity development is similar to that reported by Voldseth (1985) at Flaxton Field to the north in central Burke County.

DISCUSSION

Work by Anderson *et al.* (1960) in adjacent townships in north central Burke County found hydrocarbon production from the Midale and Rival subintervals to occur underneath paleostructures shown on a Ratcliffe isopach map (Fig. 11 of Anderson *et al.*, 1960). A similar pattern is seen on a Ratcliffe isopach map constructed for southern Burke, northern Mountrail, and eastern Ward counties (Fig. 7). The oil, however, is produced from deeper subintervals in the Mission Canyon Formation not from the Midale or Rival subintervals.

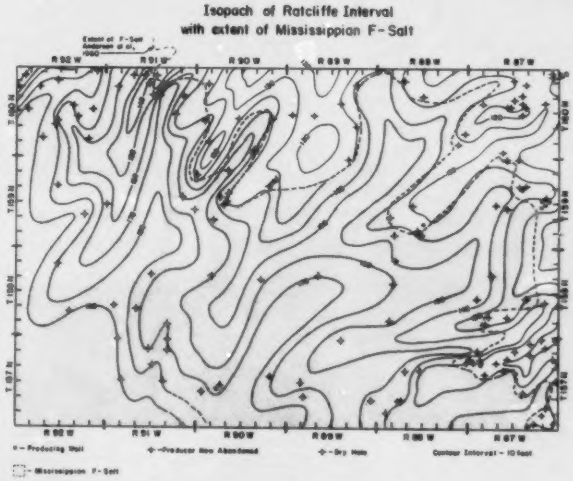
Isopach of Ratcliffe Interval
with extent of Mississippian F-Salt
Extent of F-Salt Anderson et al., 1960
R 92 W
R 91 W
R 90 W
R 89 W
R 88 W
R 87 W
T 160 N
T 159 N
T 158 N
T 157 N
Producing Well
Producer Now Abandoned
Dry Hole
Contour Interval - 10 feet
Mississippian F-Salt

Figure 8. Map of the F salt in the stippled area, with the zero edge indicated by the dashed line. Note coincidence of producing wells with the edge of the F salt.

When a map of the F salt is superimposed on the Ratcliffe isopach map (Fig. 8), one sees that oil production occurs along the edge of the overlying salt. The production here is most frequently from the Coteau subinterval. An isopach map of the Coteau subinterval beneath the Ratcliffe interval shows that production is from the flanks of Coteau paleostructures (Fig. 5). Unlike north central Burke County, the Midale and Rival of southern Burke, northern Mountrail, and northeastern Ward have not produced oil to date. However, the similarities of the structural and depositional setting appear identical and may indicate that hydrocarbons remain to be discovered.

Structural cross-sections of Vanville and Niobe Fields show that there is an element of structural control to trapping hydrocarbons (Fig. 6), but that other factors also may be contributing to trap formation. Permeability barriers created by anhydrite plugging of porosity were found to contribute to Midale and Rival hydrocarbon traps (Anderson *et al.*, 1960; Hendricks, 1988; Hendricks and Eisel, 1988), and to Coteau and Bluell (Voldseth, 1987) traps in north central Burke County. It is very likely that similar permeability barriers are contributing to trap development in southern Burke County and the surrounding area as well. This may explain why the structurally higher Caston-State No. 2 in Vanville Field had less production than the down-dip Castor State No. 1 (Fig. 6a).

CONCLUSIONS

A technique learned during early exploration for shallower Charles Formation production has application to exploration for oil in the deeper Mission Canyon Formation. The technique involves mapping the edge of the F-salt subinterval of the Ratcliffe interval. The zero isocline outlines thickness changes that imply paleostructures where hydrocarbons have accumulated. Additional structure and isopach maps of the subinterval of interest are needed for prospecting, but the F salt edge map provides a quick illustration of prospective areas for more detailed consideration.

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EXPLOITATION OF A MISSSISSIPPIAN CARBONATE RESERVOIR THROUGH HORIZONTAL DRILLING — THE GAINSBOROUGH NORTH POOL

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ABSTRACT

The Gainsborough North Pool provides an example of maximizing the production from a reservoir by utilizing horizontal drilling. The discovery well, drilled in December, 1987, encountered an oil saturated carbonate reservoir rock within the Mississippian Frobisher, Kisbey and Alida sequences with an estimated oil column of 23 m (75 ft). The reservoir consists of pisolitic to oolitic limestones ranging from a grainstone to a packstone. The Frobisher and Kisbey carbonates are interbedded with mudstones and silty sandstones, whereas the Alida reservoir section is more homogeneous.

Subsequent development drilling resulted in 25 vertical wells, establishing pool boundaries and additional reservoir data. This drilling revealed a pool area of approximately 1024 Ha (2560 acres), and showed the pool to have an active aquifer system, as well as fractures in portions of the reservoir. The combination of the active aquifer system, fracturing and near wellbore coning provided early water production in nearly all wells. Production at this time averaged 191 m^3 OPD (1200 BOPD) from the 19 productive vertical wells.

A horizontal drilling program was undertaken as means to enhance productivity through additional volumes, and beneficially to influence ultimate pool recovery. The first horizontal well was drilled in late 1989 and was completed for a production rate of 159 m^3 OPD (1000 BOPD). The initial success has resulted in an additional seven horizontal wells being drilled in the pool. The Gainsborough North pool is currently producing 731 m^3 OPD (4600 BOPD) from seven horizontal wells and 18 vertical wells, with a current cumulative oil production of 460,000 m^3 oil (2.9 mmstb).

TRANSGRESSIVE CYCLES IN AN OVERALL SHALLOWING UPWARDS SEQUENCE, MISSISSIPPIAN, MISSION CANYON, NOTTINGHAM UNIT, WILLISTON BASIN, SOUTHEAST SASKATCHEWAN

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ABSTRACT

The Nottingham North Alida Beds Unit is located on the northern margin of the Williston Basin in southeast Saskatchewan. Lower Mississippian shallow shelf carbonates of the Mission Canyon Formation display transgressive-regressive cycles in an overall shallowing upwards sequence. Cyclic beds average 32 m in thickness and are laterally continuous across the unit.

A cycle includes three facies in ascending order: (1) Unconformity-subaerial exposure facies (MC-2, MC-4) displaying a clay-rich and dolomitized weathered zone, (2) Intertidal Oncolite Boundstone Algal facies (Algal of MC-3, Lower MC-5) - early marine cementation accounts for preservation of primary interparticle porosity, (3) Normal Marine facies (MC-3, Upper MC-5) containing only secondary non-fabric selective leached porosity. The facies include crinoid ossicles, brachiopods, foraminifera, ostracodes, calcareous algae, solitary and colonial corals, and bryozoa.

Two fluid systems are present within each cycle: a) a thin finely laminated mudstone (often dolomitized) trapping fluids within the Intertidal Oncolite Boundstone (facies 2); and b) the Unconformity weathered zone (facies 1) trapping fluids in the Normal Marine facies (facies 3).

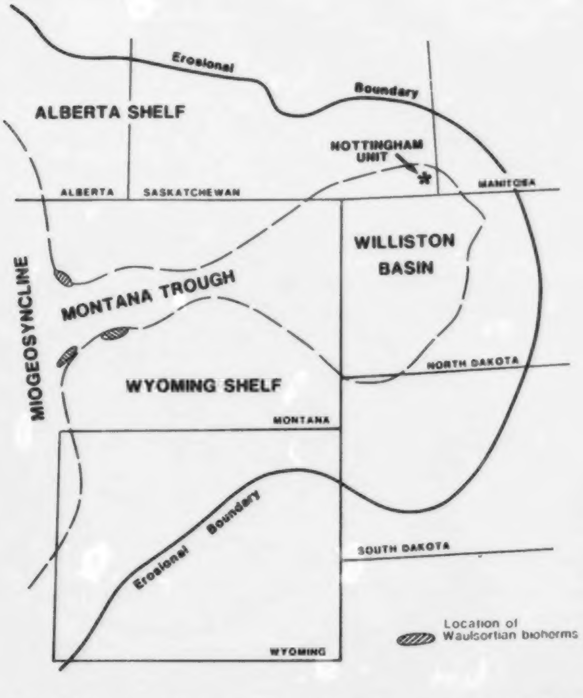


Figure 1: Location of Nottingham Unit at the northern margin of the Williston Basin. Open marine circulation was via the Montana Trough. After Bjorlie and Anderson (1978).

INTRODUCTION

The Williston Basin occupied a large area of the western North American craton during the Mississippian (Fig. 1, after Bjorlie and Anderson, 1978). It received open marine circulation from the Pacific Ocean via the Montana Trough. Mississippian carbonates were deposited in a tectonically stable environment. Post-Mississippian erosion removed progressively more section eastward and toward the margins of the Basin.

Figure 2 outlines the members of the Mississippian stratigraphic column for southeastern Saskatchewan. The entire Mississippian section shallows upwards from the basal Lodgepole/Souris Valley toe-of-slope "Waulsortian-type" mounds of the Montana trough (Cotter, 1965; Sereda and Kent 1987; Smith, 1972) to the sabkha evaporites of the Poplar Beds.

The Lodgepole/Souris Valley Beds are recognized as deep water marine carbonates (Smith, 1972) and the upper member of the Tilston Beds (MC-1) as a cross-bedded oolitic shoal. The Alida-Frobisher-Midale Beds contain deepening upwards (transgressive) cycles in an overall shallowing upward sequence (Lake, 1989).

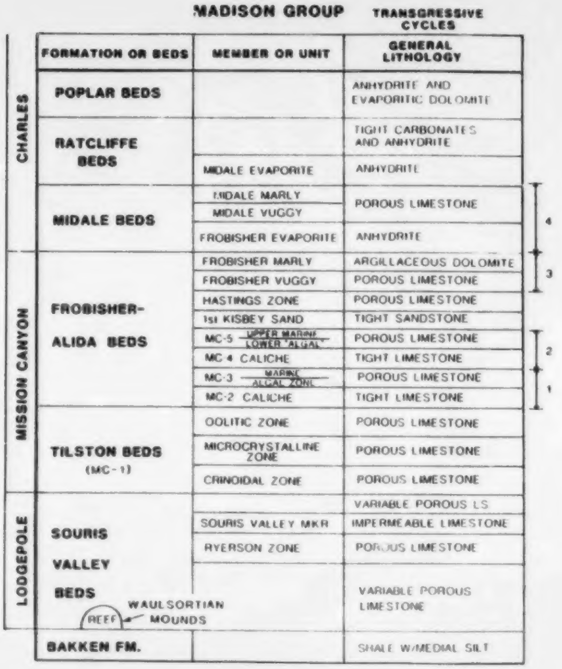
Four transgressive cycles are recognized in the shallow shelf carbonates of the Alida, Frobisher and Midale Beds (Fig. 2). This paper deals with the first two cycles (1 and 2) as preserved at Nottingham.

NATURE OF TRANSGRESSIVE CYCLES WITHIN THE MISSION CANYON

The cycles within the Alida Beds at Nottingham Unit are subdivided into three distinct facies in ascending order: 1) Unconformity- subaerial exposure facies (MC-2; MC-4); 2) Intertidal Oncolite Boundstone Algal facies (Algal of MC-3, Lower MC-5); and 3) Normal Marine facies (MC-3, Upper MC-5). Each cycle averages 32 m in thickness.

UNCONFORMITY- SUBAERIAL EXPOSURE FACIES (MC-2, MC-4)

This facies is composed of green terrigenous clays, plus dolomitized, mottled and brecciated lime mudstone interpreted to represent a caliche weathering profile. The mottled and brecciated appearance is attributed to variation in degree of weathering. The



MADISON GROUP

	FORMATION OR BEDS	MEMBER OR UNIT	GENERAL LITHOLOGY	TRANSGRESSIVE CYCLES
CHARLES	POPLAR BEDS		ANHYDRITE AND EVAPORITIC DOLOMITE	
	RATCLIFFE BEDS		TIGHT CARBONATES AND ANHYDRITE	
		MIDALE EVAPORITE	ANHYDRITE	
	MIDALE BEDS	MIDALE MARLY MIDALE VUGGY	POROUS LIMESTONE	4
		FROBISHER EVAPORITE	ANHYDRITE	
MISSION CANYON	FROBISHER-ALIDA BEDS	FROBISHER MARLY	ARGILLACEOUS DOLOMITE	3
		FROBISHER VUGGY	POROUS LIMESTONE	
		HASTINGS ZONE	POROUS LIMESTONE	
		1st KISBEY SAND	TIGHT SANDSTONE	
		MC-5 UPPER MARINE LOWER "ALGAL"	POROUS LIMESTONE	2
		MC-4 CALICHE	TIGHT LIMESTONE	
		MC-3 MARINE ALGAL ZONE	POROUS LIMESTONE	1
		MC-2 CALICHE	TIGHT LIMESTONE	
	TILSTON BEDS (MC-1)	OOLITIC ZONE	POROUS LIMESTONE	
		MICROCRYSTALLINE ZONE	POROUS LIMESTONE	
		CRINOIDAL ZONE	POROUS LIMESTONE	
LODGEPOLE	SOURIS VALLEY BEDS		VARIABLE POROUS LS	
		SOURIS VALLEY MKR	IMPERMEABLE LIMESTONE	
		RYERSON ZONE	POROUS LIMESTONE	
			VARIABLE POROUS LIMESTONE	
	BAKKEN FM.		SHALE W/MEDIAL SILT	

Figure 2: Mississippian stratigraphic column for southeast Saskatchewan. Note the four transgressive cycles within the Mission Canyon and Lower Charles.

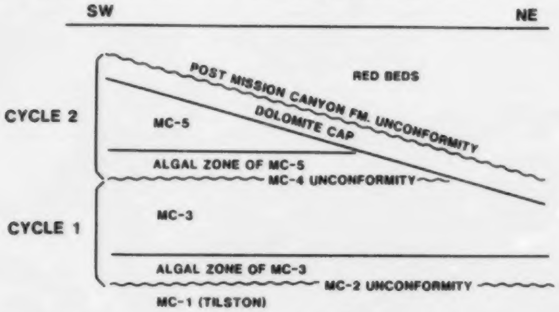


Figure 3: Idealized cross-section displaying two transgressive Mission Canyon cycles at Nottingham.

green clays occur as 1) several centimetres of laminated shale, 2) infill of subvertical weathering fractures and 3) disseminated clays throughout the dolomitized lime mudstone. The gamma ray log indicates a radioactive marker bed. Well-preserved plant remains were observed immediately below the green shale in two cores. The excellent preservation is attributed to a low energy environment of deposition.

INTERTIDAL ONCOLITE ALGAL FACIES (ALGAL ZONE OF MC-3; LOWER MC-5)

This facies is composed of numerous thin lenses of oncolites, each grading into continuous algal coatings, terminated by 1 mm-thick clay laminae. Flugel (1982) would have oncolites originate in lower energy agitated waters than oolites. Oncolites lack foreign nucleii but like oolites, they are algal spheres which grew by agglutinating particles to their mucous surface. The oncolites show evidence of storm aggradation, with many grains being nucleated around broken fragments. The Algal facies is devoid of well-preserved invertebrate remains.

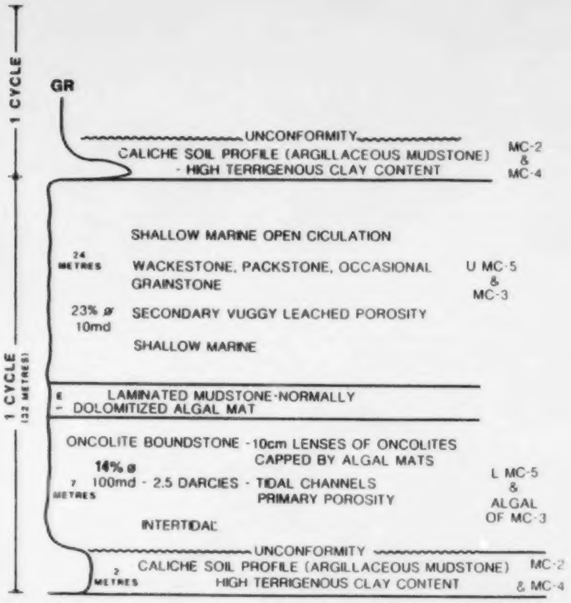

1 CYCLE
GR
UNCONFORMITY
CALICHE SOIL PROFILE (ARGILLACEOUS MUDSTONE)
- HIGH TERRIGENOUS CLAY CONTENT
MC-2 & MC-4
1 CYCLE (32 METRES)
24 METRES
SHALLOW MARINE OPEN CICULATION
WACKESTONE, PACKSTONE, OCCASIONAL GRAINSTONE
U MC-5 & MC-3
23% ø 10md
SECONDARY VUGGY LEACHED POROSITY
SHALLOW MARINE
LAMINATED MUDSTONE-NORMALLY
DOLOMITIZED ALGAL MAT
ONCOLITE BOUNDSTONE -10cm LENSES OF ONCOLITES CAPPED BY ALGAL MATS
14% ø
7 METRES
100md - 2.5 DARCIES - TIDAL CHANNELS PRIMARY POROSITY
L MC-5 & ALGAL OF MC-3
INTERTIDAL
UNCONFORMITY
2 METRES
CALICHE SOIL PROFILE (ARGILLACEOUS MUDSTONE)
HIGH TERRIGENOUS CLAY CONTENT
MC-2 & MC-4

Figure 4: Idealized Mississippian transgressive cycle at Nottingham.

The facies is interpreted as intertidal and hypersaline based on the thin nature of the lenses, absence of other fauna and presence of thin hardgrounds. The seven-metre average thickness of the Algal zone is evident from figure 5. A tidal channel is oriented east-west, parallel to the basin margin (Fig. 6). It is recognized by a thinning of the algal interval, and is infilled with mud and thin impermeable oncolite lenses to the thickness of the Algal isopach. The Algal facies is capped by a finely laminated mudstone which is often dolomitized but sedimentary texture is preserved.

NORMAL MARINE FACIES (MC-3, UPPER MC-5)

Normal Marine facies is composed of wackestone and packstone containing crinoid ossicles, brachiopods, foraminifera, ostracodes, calcispheres, solitary and colonial corals, bryozoa and tubular dasycladian algae. Its environment of deposition is interpreted as low energy shallow marine as based on the diversity of fauna and abundance of carbonate mud. The allochems display varying degrees of abrasion. Early compaction is evidenced by broken brachiopod shells and deformed dasycladian tubular algae.

DIAGENESIS WITHIN THE MISSION CANYON FORMATION

The three facies represented in each deepening-upwards cycle have undergone diagenetic alteration but remain largely undolomitized.

UNCONFORMITY/SUBAERIAL EXPOSURE FACIES (MC-2; MC-4)

The unconformity surfaces have been altered by subaerial weathering and primary textures have been largely obliterated. Typical caliche texture with indistinct boundaries for unweathered clasts and vertical fractures infilled with green clay are displayed. Primary pores have been infilled and this facies acts as a reservoir

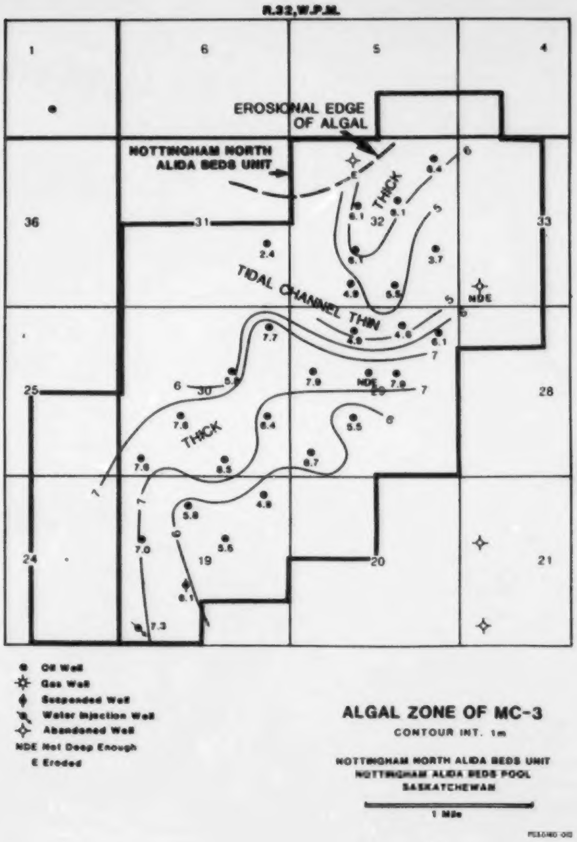


Figure 5: Isopach Map, Algal zone of the MC-3 at Nottingham. Note the east-west trending tidal channel represented by the thinning of the Algal zone.

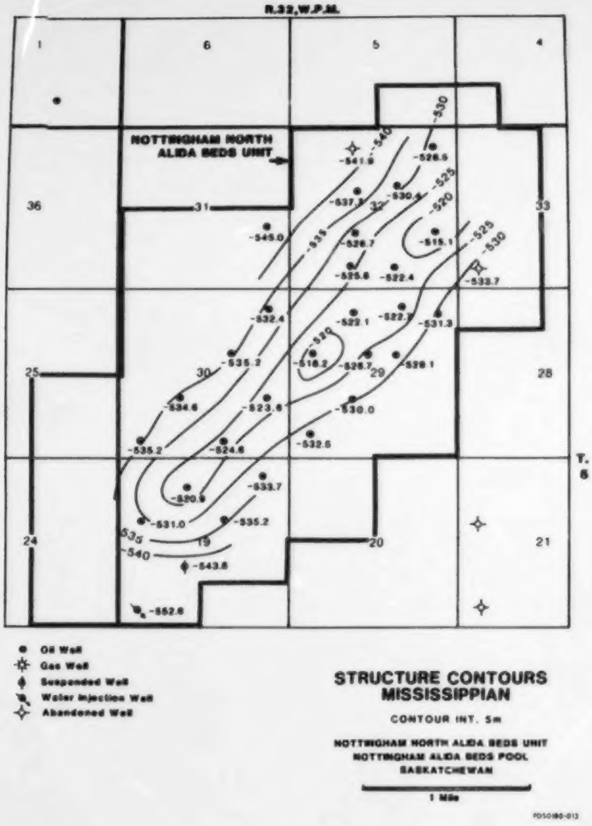

R.32,W.P.M.
NOTTINGHAM NORTH ALIDA BEDS UNIT
T. 8
Oil Well
Gas Well
Suspended Well
Water Injection Well
Abandoned Well
STRUCTURE CONTOURS MISSISSIPPIAN
CONTOUR INT. 5m
NOTTINGHAM NORTH ALIDA BEDS UNIT
NOTTINGHAM ALIDA BEDS POOL
SASKATCHEWAN
1 Mile

Figure 6: Structure contours, top of the Mississippian at Nottingham.

cap to the underlying Normal Marine facies. The weathering profile displays a gradual increase in green clay content towards the unconformity surface as reflected in the gamma ray response.

INTERTIDAL ONCOLITE ALGAL FACIES

The oncolite boundstones of the intertidal algal zones contain coated isopachous bladed crystals, interpreted as early marine. The oncolites do not show evidence of compaction or pressure solution at grain to grain contacts, suggesting early marine cementation. The preservation of primary porosity is high, resulting in the excellent permeabilities measured in these rocks. Porosity reduction occurs as a result of diagenetic growth of anhydrite and siderite near the base of some algal zones. Kendall and Walters (1978) suggest that the anhydrite was formed at the time of oil migration. The iron for the siderite may have come from the underlying unconformity/subaerial exposure facies.

NORMAL MARINE FACIES

The Normal Marine sequences display a severe reduction in porosity through extensive carbonate cementation and late-stage anhydrite infill. Syntaxial overgrowths of sparry calcite have eliminated primary porosity which likely was limited because of the muddy nature of the matrix. Only secondary vuggy or leached porosity is developed and that is fabric-selective. Dasyclad algae, crinoid ossicles, brachiopods and matrix lime mud are all leached.

DOLOMITE CAPROCK

The Mississippian system in most wells of the Nottingham Unit is capped by a low permeability dolomite which averages 6 m in thickness. Since the dolomite cap cross-cuts stratigraphy present at the subcrop, it is considered to be late stage diagenetic. The cap is a rhombic sparry replacement of dolomite which has preserved the primary texture. Large crinoidal allochems are preserved as calcitic forms or as anhydrite pseudomorphs. The contact between dolomite cap and underlying limestone is abrupt. The dolomite contains some porosity but is impermeable to fluid flow; hence it acts as a seal for the upper reservoir units of Nottingham Unit.

Post-burial dewatering of the overlying clay-rich Jurassic Lower Watrous Red Beds is suggested as the source of magnesium for dolomitization of the Mississippian subcrop. The Upper Watrous sabkha evaporites/dolomites prevented upward escape of meteoric waters from the red beds; thus dolomitizing magnesium-rich waters were forced downwards into the porous carbonates.

Plate 1: Unconformity-subaerial exposure facies; (MC-2, MC-4) caliche (soil profile) with high terrigenous clay representing subaerial exposure at top of MC-2. Well 14-32-5-32 WPM: 3710', MC-2.

Plate 2: Intertidal Oncolite Algal facies. Core photograph of Algal zone reservoir rock with preservation of primary porosity between grains. Thin algal mats periodically encrust the oncolites and represent a reduction in energy conditions. Well 10-29-5-32 WPM: 3694', Algal zone of MC-3.

Plate 3: Intertidal Oncolite Algal facies. Photomicrograph of oncolites encrusted by algal mat. The oncolites are truncated by a thin hardground and encrusted by algae in response to a reduction in water agitation. Shallowing upwards sequences on the order of tens of centimetres are recognized throughout the Algal zone. Well 10-32-5-32 WPM: 3667', PPL 2.5X, Algal zone of MC-3; porosity= 20.9 percent, permeability= 70 md.

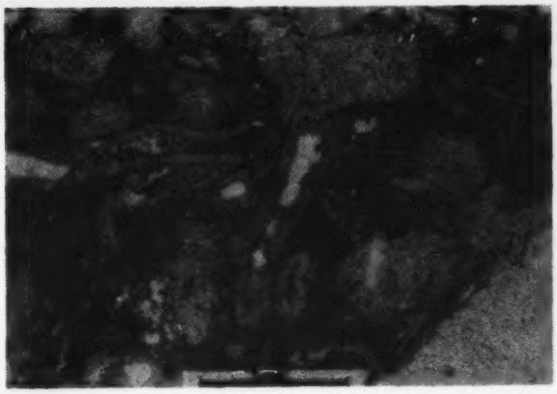


Plate 4: Normal Marine facies (MC-3, Upper MC-5). Photomicrograph of porous crinoidal packstone (MC-3) reservoir rock. Thin section depicts a diversity of abraded marine fossils including crinoids and calcareous algae. Secondary porosity is developed as intra-skeletal or intra-particle voids within calcareous algae. Well 10-30-5-32 WFM: 3712', ppl 2.5x, porosity= 11.1 percent, permeability= 2.1 md.

Plate 5: Dolomite Caprock. Photograph of core displaying top of Mississippian section. The left half of the photo shows the tight dolomite caprock (white). In contrast, the porous and permeable oil-stained MC-3 reservoir limestone is on the right. The dolomite caprock traps hydrocarbons. Lithology is identical across this boundary. Well 10-29-5-32 WPM: 3682.5-3707.5', Dolomite Cap.

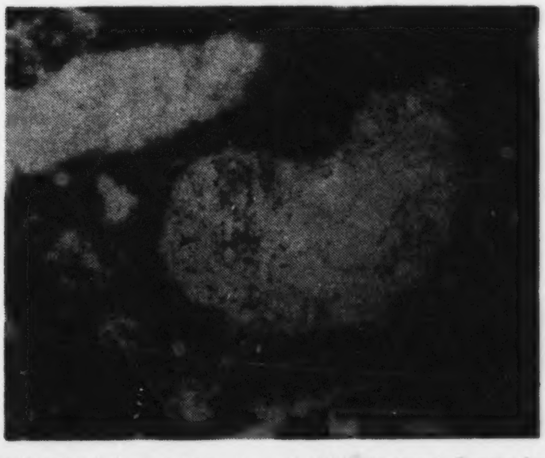


Plate 6: Dolomite Caprock. Photomicrograph of dolomite caprock. Remnant fossil ghosts have been preserved as calcite or replaced by anhydrite. The matrix material of the original limestone was dolomitized first. Large allochems are preserved as calcite or subsequently replaced by anhydrite. Large allochems are preserved as calcite or anhydrite. Well 6-32-5-32 WPM: 3644', ppl 6.3x, Dolomite Cap.

RESERVOIR QUALITY

The Algal zones of the MC-3 and Lower MC-5 have the best reservoir quality with average porosity and permeability values of 14 percent and 100 millidarcies respectively. Permeability values to 2.5 darcies have been measured and are attributed to preservation of primary interparticle porosity. In contrast, the Normal Marine sequences of the MC-3 and Upper MC-5 with their secondary leached vuggy pores average 17 percent porosity and 10 millidarcies permeability.

There are four separate fluid systems, each with separate fluid contacts: Algal zone of MC-3, MC-3, Lower MC-5, and Upper MC-5 (Fig. 1). The systems are separated by the impermeable dolomitic mudstones at the unconformities (MC-2 and MC-4) and tight, finely laminated dolomitic mudstone of the Algal zone (MC-3) and Lower MC-5. The cyclic nature of the sediments resulted in vertically stacked reservoirs. Each fluid system has an active water drive which is part of the regional hydrodynamic system and sourced at the western margin of the Williston Basin in western Montana. Waterflooding is not necessary as pressure is maintained by the natural water drive.

POST-MISSISSIPPIAN EROSION

The stratigraphic section from the Mississippian to the Lower Jurassic, corresponding to a period in excess of 100 million years, is missing in the Nottingham Unit. Mississippian strata are successively eroded to lower stratigraphic levels northeasterly towards the margin of the basin. A complex topography was developed on the erosional surface. Hydrocarbons are trapped at the updip truncational edge of the reservoir units. Paleotopographic relief of the Mississippian highs containing the oil reservoirs is in the order of 35 m. Pools in southeastern Saskatchewan are separated from one another by erosional valleys (trending N 30° E) which are infilled with Jurassic Red Bed/Lower Watrous sands. The orientation of these valleys is that of the regional fracture system. However, infill of secondary channels, perpendicular to the long axes of the pools, here and there isolate the reservoir from the active water drive.

SUMMARY

This study of the Mission Canyon Formation identified two deepening upwards sedimentation cycles at Nottingham Unit, each averaging 32 m in thickness. During Mississippian time in the Williston Basin, an overall shallowing upwards sequence containing several transgressive cycles was deposited. Each transgressive cycle at Nottingham begins with a subaerial exposure and a caliche soil profile (MC-2, MC-4), which is overlain by intertidal oncolite boundstone with primary interparticle porosity. The boundstone was deposited in response to a marine incursion (Algal zone of MC-3, Lower MC-5). Tidal channels within this facies, depicted by boundstone isopach thins, have an east-west orientation that is parallel to the shelf margin. The oncolite boundstone is capped by a one-metre laminated mudstone which is often dolomitized. This impermeable bed isolates the boundstone reservoir from the uppermost unit of the cycles. The laminated mudstone represents a short pulse of shallowing in the overall deepening upwards cycle. The oncolite boundstone is overlain by a shallow marine sequence containing a diverse invertebrate fauna (MC-3, Upper MC-5). Primary porosity in the upper unit was occluded by cementation after burial and later modified through a change in water chemistry to produce secondary vuggy or leached porosity with low permeability.

In the hiatus from Mississippian to Lower Jurassic (a period in excess of 100 million years) a complex topography controlled by the regional fracture system was formed. Magnesium-rich water in the overlying Jurassic Red Bed sands could not move upwards after burial and consequently entered and dolomitized the subjacent Mississippian strata. This impermeable dolomite caprock averages 6 m in thickness and forms the uppermost reservoir seal at Nottingham.

REGIONAL EXTRAPOLATION

The Frobisher and Midale Beds are made up of transgressive cycles similar to the two cycles within the Alida beds at Nottingham. This makes a total of four transgressive cycles preserved within the Mississippian of the Williston Basin: (1) Alida MC-2, Algal, MC-3; (2) Alida MC-4, Lower and Upper MC-5; (3) Frobisher Vuggy and Marly; and (4) Frobisher Evaporite, Midale Vuggy and Marly.

The Vuggy members of the Frobisher and Midale Beds correspond to the intertidal oncolite boundstones of the Alida transgressive cycles. Dolomitization at Weyburn, Midale and Steelman pools has partly obscured the primary texture; however, the superior nature of the porosity and permeability of this facies is still evident.

The Marly member of the Frobisher and Midale Beds has been altered to a microcrystalline dolomite with intercrystalline porosity and low permeability. The Marly corresponds to the Normal Marine facies of the Alida Beds. Original textures are obscured but the relationships are the same.

Waters and Sando (1978) and Hoff (1987) recognized transgressive cycles within the Mission Canyon Formation of the Williston Basin. Uniformly shallow water depth over a great lateral area is demonstrated by the regional correlation of units around the basin margin.

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RESERVOIR STYLES AND POROSITY CHARACTERISTICS OF THE MISSISSIPPIAN RATCLIFFE BEDS: OUNGRE TO LAKE ALMA FIELD TRANSECT, SOUTHEASTERN SASKATCHEWAN

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ABSTRACT

A number of reservoir styles, producing lithologies, and associated porosity systems have been identified in the Mississippian Ratcliffe Beds along a transect from the Oungre Field, through the Flat Lake Field, to the Lake Alma Field. Scanning electron microscopy (SEM) and mercury- injection capillary pressure analysis (MICP) are important techniques in studying these carbonate rocks.

At the Oungre Field (Township 2, Range 14W2M) a stratigraphic trap is created within the Oungre zone by the change from non-argillaceous microsucrosic dolomite to argillaceous carbonate up-dip. Porosity is homogeneous, very fine intercrystalline, with minor remnant micrite as observed by SEM. MICP analyses show a high degree of uniformity in pore throat sizes, and high ejection efficiencies.

In the Flat Lake Field (Township 1, Range 16W2M) oil is trapped in the upper Ratcliffe Beds by a combination of structural and stratigraphic elements. Reservoir lithologies include oolitic-pelletoidal-skeletal wackestones to packstones, lime mudstones, and dolomitic mudstones. The wackestone-packstone lithofacies developed as low-relief shoals near the edge of the Hummingbird Trough. Intergranular and intercrystalline porosity types are present in reservoir strata. MICP analyses indicate heterogeneous pore networks in Flat Lake reservoir rocks, though ejection efficiencies are comparable to those at the Oungre Field.

In the Lake Alma Field (Township 1, Range 17W2M) oil is trapped by structural closure (over 40 m on the Ratcliffe Beds) related to dissolution of salt in the Prairie Evaporite (Devonian). As in the Oungre Field, excellent intercrystalline porosity is developed in microsucrosic dolomites of the Oungre zone. In addition, microcrystalline limestones serve as reservoir rock to a much lesser extent. Microvugs and moldic porosity cause greater heterogeneity of pore systems, though ejection efficiencies remain high.

The quality of these Ratcliffe Beds reservoirs is controlled by depositional factors which influence the extent of dolomitization.

INTRODUCTION

In southeastern Saskatchewan, the Mississippian Ratcliffe Beds host hydrocarbons in a variety of reservoir styles and porosity systems. Production from this interval, on the Canadian side of the Williston Basin, is confined to the area within Townships 1 to 5, Ranges 13 to 19W2M (Fig. 1). Three types of reservoir/trap styles are seen in the Ratcliffe Beds: stratigraphic pools of the 'Oungre Trend', structural pools such as Lake Alma and Hummingbird fields, and stratigraphic - structural pools such as Flat Lake.

The study of three representative pools, Oungre, Flat Lake, and Lake Alma, along a northeast to southwest transect involved: 1) core study to develop a depositional model, and 2) investigation of the pore systems of reservoir lithologies, using mercury-injection capillary pressure techniques (MICP) supplemented by scanning electron microscopy (SEM). Fifty cores from the three fields were examined and 60 MICP samples were analyzed.

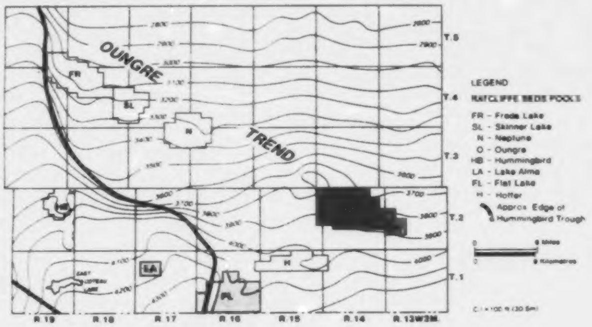


Figure 1. Distribution of Ratcliffe Beds pools in southern Saskatchewan; structure contours on the Ratcliffe Beds. Stippled regions are fields included in this study. (Modified after Kent *et al.*, 1988)

GEOLOGICAL SETTING

Mississippian strata of southern Saskatchewan lie on the northern flank of the Williston Basin, in an intracratonic sedimentary basin centred near Williston, North Dakota (Fig. 2) (Fuzesy, 1973). The Ratcliffe Beds lie conformably between the Poplar Beds above, and the Midale Beds below. These three units combined are equivalent to the Charles Formation (Fig. 3). This sequence is composed of marker-defined carbonates and evaporites deposited as 'brining-upward' cycles (Longman *et al.*, 1983) during overall basin infilling. In the study area, the first cycle of the Ratcliffe Beds begins with deposition of the Oungre zone carbonate and culminates with the Oungre Evaporite; the second cycle records upper Ratcliffe zone carbonate capped by the upper Ratcliffe evaporite.

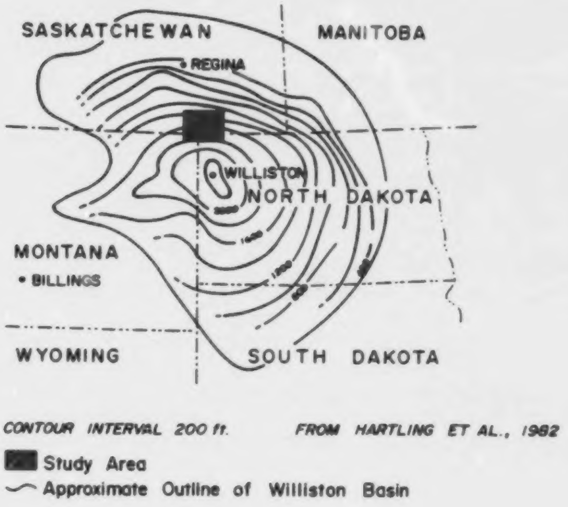


Figure 2. Approximate position of study area in the Williston Basin (from Hartling *et al.*, 1982).

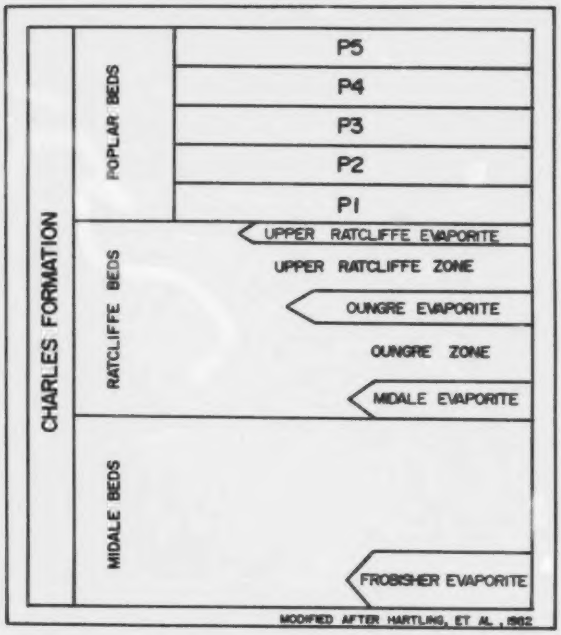


Figure 3. Mississippian Charles Formation nomenclature in southeastern Saskatchewan.

SEDIMENTOLOGY AND DEPOSITIONAL ENVIRONMENTS

Within the study area the Ratcliffe Beds contain two potential carbonate reservoir units. The lower, the Oungre zone, is separated from the upper Ratcliffe zone by the Oungre Evaporite (Fig. 3).

The Oungre zone is characterized by porous zones of microsucrosic dolomite or dolomitic limestone, interbedded with non-porous units of cryptocrystalline or anhydritic dolomite, dense lime mudstone/wackestone, or calcite- or anhydrite-cemented oolitic-pelletoidal-skeletal packstone/grainstone. The volume of packstone/grainstone lithology increases from the Oungre to Lake Alma field, reflecting a progression to higher energy conditions at the latter.

The best reservoir dolomites are commonly found within a few metres of the Oungre Evaporite base, though at Lake Alma they are also found deeper in the Oungre zone.

Lime muds of the Oungre zone, many of which were subsequently dolomitized to some degree, were deposited in four environments as defined by Hartling *et al.* (1982): 1) shoals (banks); 2) intershoal regions; 3) tidal channels; and 4) lagoons. In a regional sense, deposition of lime muds took place in an open to slightly restricted, low energy, shallow shelf setting. Evaporitic conditions existed locally.

The oolitic-pelletoidal-skeletal sediments were derived from the shelf margin regions farther basinward where they formed low relief shoals. Storm surge periodically washed these sediments leeward into the shelf interior (Hartling *et al.*, 1982).

The upper zone of the Ratcliffe Beds is more heterogeneous. In the vicinity of the Flat Lake Field, oolitic-pelletoidal-skeletal wackestones to packstones occur interbedded with dolomitic or lime mudstones, or micrite-rich dolomites. Elsewhere, chalky lime mudstones, or argillaceous, bioturbated wackestones are common. A few evaporitic carbonate and nodular anhydrite beds may also be present.

The depositional setting for the upper Ratcliffe zone was an open to semi-restricted shallow shelf, similar to that of the Oungre zone, although locally, higher energy, shoaling conditions existed, *e.g.*, at Flat Lake pool. Chalky carbonates are the product of a lagoonal environment (Edie, 1958), where lower energy, more restricted conditions prevailed.

The Oungre Evaporite is mainly massive to bedded standing water anhydrite. Precipitation of these evaporites occurred in hypersaline restricted lagoons behind a barrier shoal complex (Warren and Kendall, 1985). Nodular and mosaic anhydrite present in some parts of the unit developed within intertidal to supratidal carbonate sediments.

OUNGRE FIELD (TOWNSHIP 2, RANGE 14W2M)

The Oungre Field is one of four pools in a northwest-southeast fairway of production referred to as the Oungre Trend (Fig. 1). These pools are largely stratigraphic in nature and all produce oil from the Oungre zone.

Dolomitization is recognized as the key element in creating reservoir along the entire trend (Hartling, 1984; Hartling *et al.*, 1982; Fuzesy, 1973). A seepage refluxion mechanism similar to that proposed by Adams and Rhodes (1960) is favoured as the cause of most of these early diagenetic dolomites (Burton, 1990; Hartling *et al.*, 1982) (Fig. 4). Reservoir dolomites of the Oungre zone at Oungre Field are typically microcrystalline (4 to 8 micrometres) to microsucrosic (8 to 14 micrometres) and have attained inhibition boundary growth stage (Kent *et al.*, 1984; Wardlaw, 1979). The fine crystallinity of these rocks is a product of the texture of the original lime sediments, and the penecontemporaneous origin of the dolomite. Uniform in appearance, allochems and moldic or vuggy porosity are rarely seen in these dolomites.

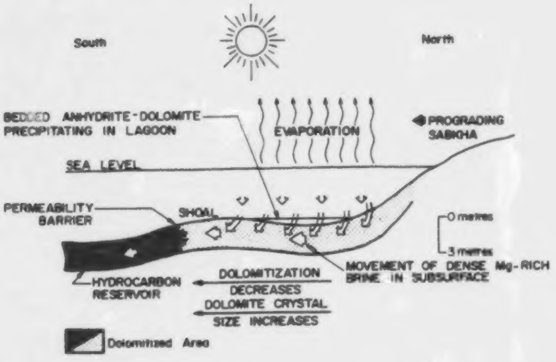


Figure 4. Model of seepage refluxion as adapted to Oungre zone sediments by Hartling *et al.* (1982).

Scanning electron microscopy (SEM) shows most of the dolomites at the Oungre pool to be remarkably free of micrite (Fig. 5). Kent *et al.*, (1984) suggested that rocks with this property may have been subjected to an episode of 'washing' by meteoric waters during a late diagenetic event.

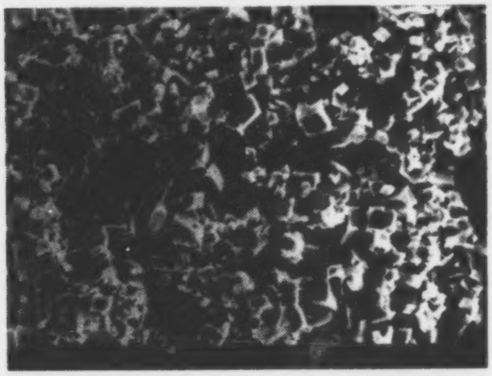


Figure 5. Scanning electron micrograph of Specimen O-11 (9-24-1-15W2M; 1869 m). 'Clean', non-micritic dolomite from the Oungre zone. Scale bar is 40 micrometres.

Porosity and permeability, and several petrophysical parameters derived from MICP analysis are useful in describing reservoir lithologies (Fig. 6). Oungre Field samples have an average porosity of 29 percent and average permeability of 13 millidarcies (Table 1). These samples are also characterized by a low standard deviation of pore throat sizes, determined indirectly from the injection profile, which attests to the high degree of uniformity in these dolomite reservoir rocks. This uniformity is also reflected in the flat to shallowly inclined plateau on most MICP curves for this field (Fig. 7). Ejection efficiencies, a measure of the ability of the reservoir to release non-wetting fluids, are excellent, typical of effectively dolomitized reservoir, and comparable to the other fields studied.

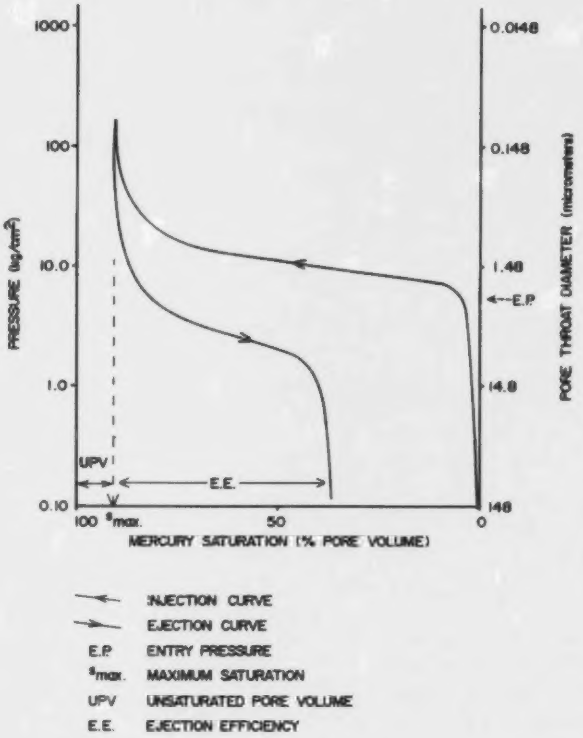
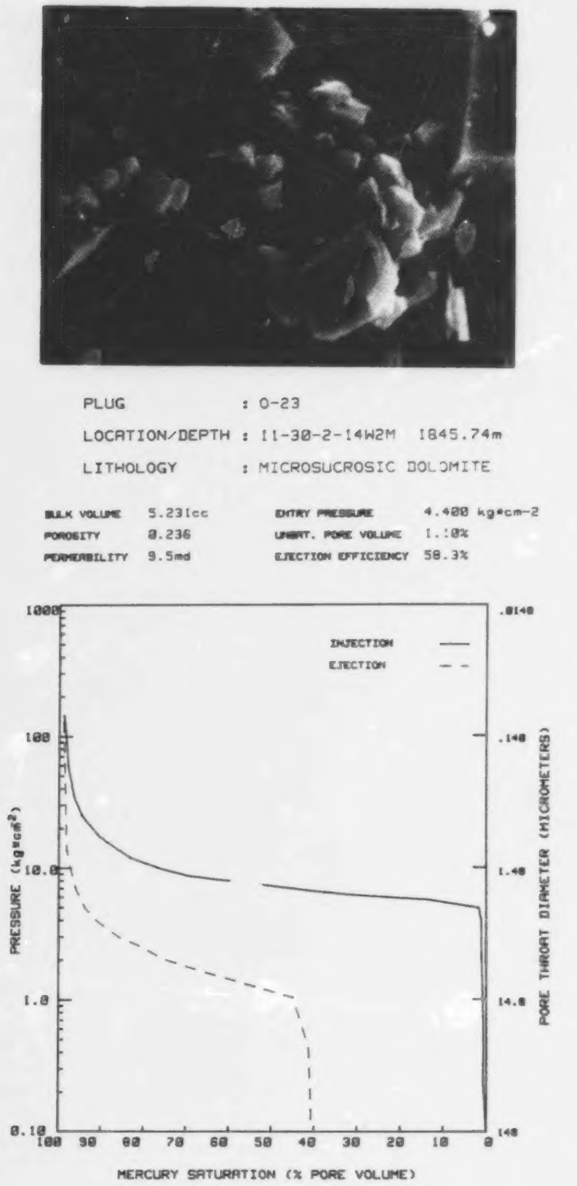


Figure 6. Petrophysical parameters determined through mercury-injection capillary pressure analysis.

Table 1: Average petrophysical values for fields.

	Oungre	Flat Lake	Lake Alma
Porosity (%)	29	18	28
Permeability (md)	12.6	7.8	12.9
Entry Pressure (kg/cm^2)	4.2	3.8	3.0
Ejection Efficiency (%)	49.7	47.0	48.1
Unsaturated Pore Volume (%)	3.5	4.2	5.5
Standard Deviation of Pore Throat Sizes	0.436	0.722	0.721

The cause of oil entrapment along the Oungre Trend is not immediately apparent. Hartling *et al.* (1982) proposed that the transition from permeable sucrosic reservoir dolomites to 'overdolomitized' impervious microcrystalline dolomite up-dip provides the necessary barrier to hydrocarbon migration. At Oungre Field another fundamental lithologic change is evident. A southwest to northeast gamma ray-sonic log cross-section through the Oungre Field reveals a change from clean Oungre zone carbonate within the pool, to argillaceous carbonate up-dip (Fig. 8). The effect of the argillaceous component is two-fold: 1) it may have lowered original, depositional permeability slightly, thus limiting movement of magnesium-rich brines and retarding the dolomitization process, and 2) more importantly, it severely restricted pore throat diameters in the very finely crystalline dolomites. These factors, in combination, have created an effective permeability barrier.



PLUG : 0-23

LOCATION/DEPTH : 11-30-2-14W2M 1845.74m

LITHOLOGY : MICROSUCROSIC DOLOMITE

BULK VOLUME 5.231cc

POROSITY 0.236

PERMEABILITY 9.5md

ENTRY PRESSURE 4.400 kg*cm-2

UNSAT. PORE VOLUME 1.10%

EJECTION EFFICIENCY 58.3%

Figure 7. Scanning electron micrograph and MICP profile of a typical dolomite reservoir rock from the Oungre Field. Specimen O-23 (11-30-2-14W2M; 1845.7 m). Scale bar is 4 micrometres.

The position of the change from effectively dolomitized reservoir to argillaceous seal is related to depositional factors. The subtle topography mapped by Hartling *et al.* (1982) and interpreted as algal-lime mud shoals (banks) and off-shoal facies at Freda Lake, Skinner Lake, and Neptune fields, is also present at Oungre Field (Figs. 9, 10). The argillaceous carbonates were deposited leeward of the mud bank complex, in the low energy, inner shelf or lagoonal setting subject to influx of fine terrigenous material.

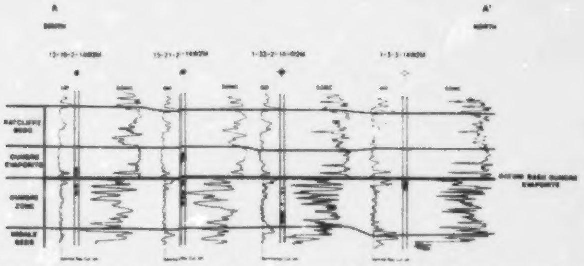


Figure 8. Stratigraphic cross-section A-A´ through the Oungre Field showing change from 'clean' reservoir grade carbonate within the pool to argillaceous, non-reservoir to the north. Gamma ray cut-off at 25 API units.

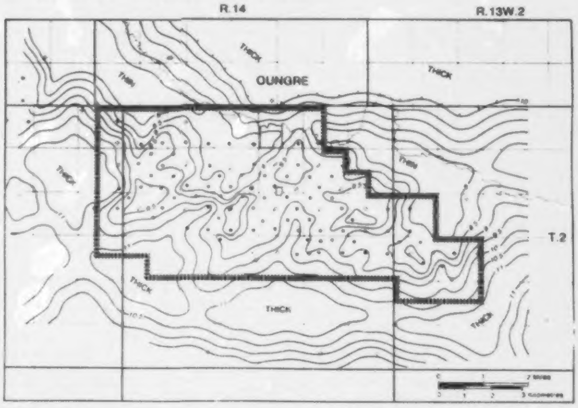


Figure 9. Isopach map of the Oungre Evaporite, Oungre Field area, reflecting depositional topography of the underlying Oungre zone carbonate interval. Contour interval : 0.5 m (1.6 ft).

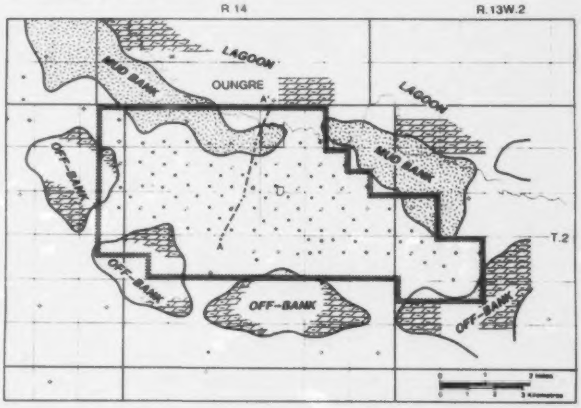


Figure 10. Paleoenvironmental map of the Oungre zone, Oungre Field area.

FLAT LAKE FIELD (TOWNSHIP 1, RANGE 16W2M)

The Flat Lake Field is situated along the eastern margin of the Hummingbird Trough, and extends southward into North Dakota (Fig. 1). Production here is from the upper Ratcliffe zone, above the Oungre Evaporite (Fig. 3). Both structural and stratigraphic elements are involved in trapping of oil in the Ratcliffe Beds. The western limit of the pool (Canadian side) is structurally controlled by the Hummingbird Trough. The eastern limit is at least partially controlled by another, less pronounced, structural low, probably related to salt dissolution. Stratigraphic controls are discussed below.

Reservoir lithologies at Flat Lake include oolitic-pelletoidal-skeletal wackestones to packstones, dolomitic lime mudstones-wackestones, and micrite-bearing dolomites.

The thick wackestone-packstone shoal facies (algal pelletoid reef facies of Hansen, 1966) was controlled indirectly by the edge of the salt trough. Marginally higher energy, shoaling conditions were favoured at the subtle bathymetric break in the Ratcliffe shelf above this lineament.

An isopach of net porosity in the upper Ratcliffe zone shows this reservoir thinning abruptly to the north and east (Fig. 11). Chalky lime mudstones are found in place of reservoir rocks in these directions. Permeabilities in the former usually drop to less than 1 mD and create an effective permeability barrier. A gamma ray - sonic log profile through the pool shows the character change from thick, wackestone-packstone reservoir facies, to thin and chalky non-reservoir facies (Fig. 12). The paleoenvironmental interpretation is presented in Figure 13.

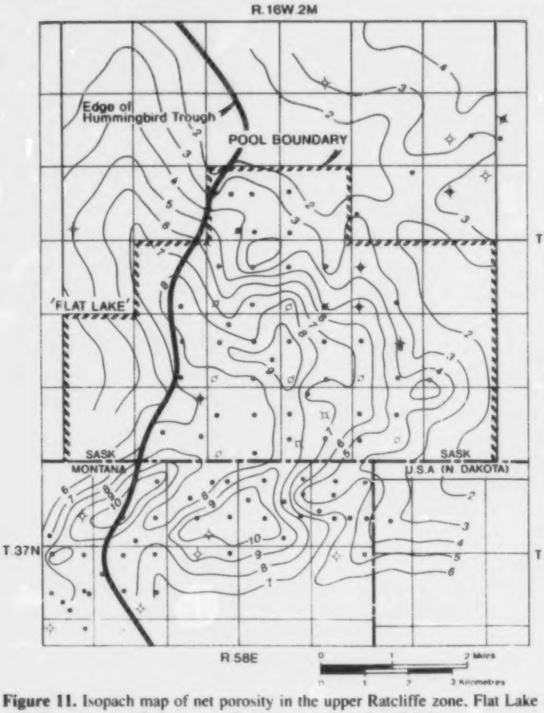


Figure 11. Isopach map of net porosity in the upper Ratcliffe zone, Flat Lake Field area. Porosity cut-off : 8 percent. Contour interval : 1.0 m (3.3 ft).

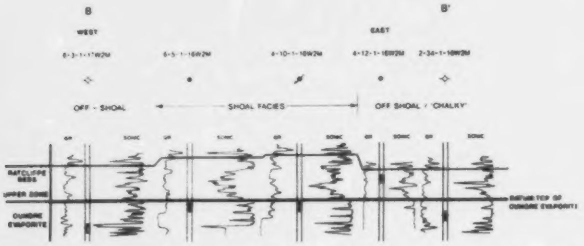


Figure 12. Stratigraphic cross-section B-B′ through the Flat Lake Field showing development of thick shoal facies (reservoir) at the pool, and change to thin, chalky (non-reservoir) facies away from the pool.

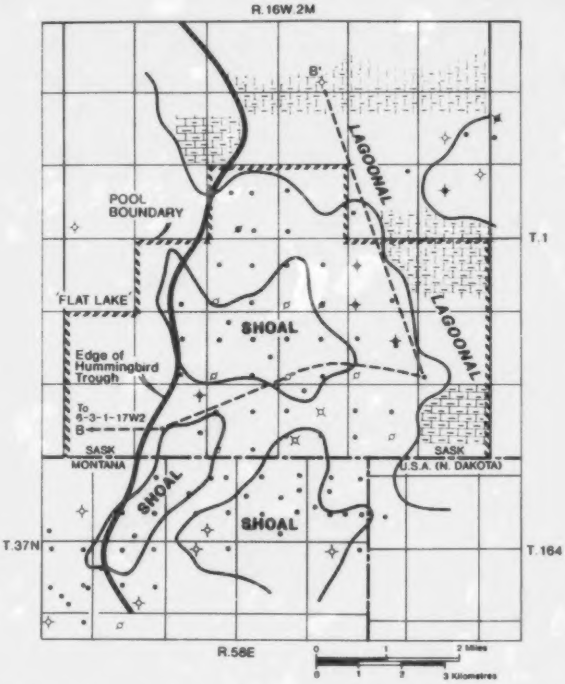


Figure 13. Paleoenvironmental map of the upper Ratcliffe zone, Flat Lake Field area.

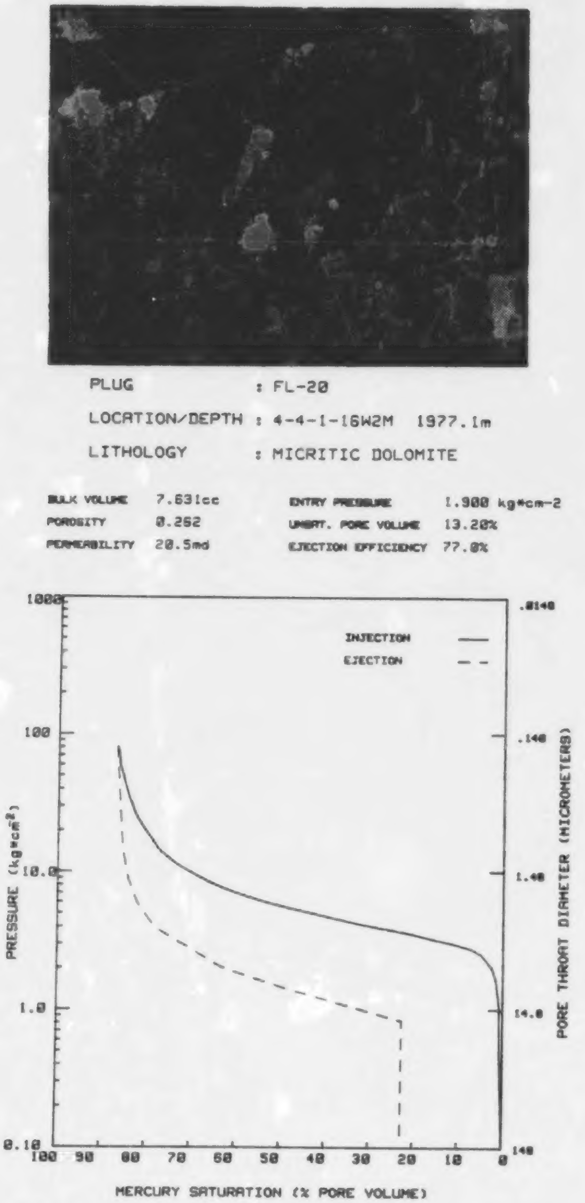
Porosity types in the reservoir rocks at Flat Lake include intercrystalline, intergranular, and microvuggy and moldic. Dolomitization, though less extensive than in the Oungre zone, remains important in creating effective intercrystalline porosity. However, the process was incomplete as these rocks often include abundant lime mud (Fig. 14).

Though rocks with some proportion of grains are abundant, few are grain-supported, and intergranular porosity is of minor importance (Fig. 15). Rarely, it is found in combination with one or more of the other porosity types.

The MICP/SEM sample population from Flat Lake reservoir rocks has a high standard deviation of pore throat sizes (Table 1). Steeper slopes on MICP profiles of many Flat Lake reservoir samples (see Figs. 13, 14) confirm the wider distribution of pore throat sizes which accompany lithologic heterogeneity. The factors responsible for this include: 1) the presence and abundance of micrite in dolomites, 2) partial dolomitization of lime sediments, and 3) combinations of intercrystalline, intergranular, or microvuggy and moldic porosity.

The pore system heterogeneity does not appear to have adversely affected ejection efficiencies, as the pool average for Flat Lake samples is comparable to that of Oungre and Lake Alma fields.

Vertical fracturing of the Ratcliffe Beds is locally intense at Flat Lake. Fractures are open, or sealed with late diagenetic, metasomatic anhydrite (Kendall and Walters, 1978). Fracturing may be related to dissolution of Devonian-aged salts at the edge of the Hummingbird Trough and collapse of overlying competent strata. The effect of fracturing on production from the pool is unknown.



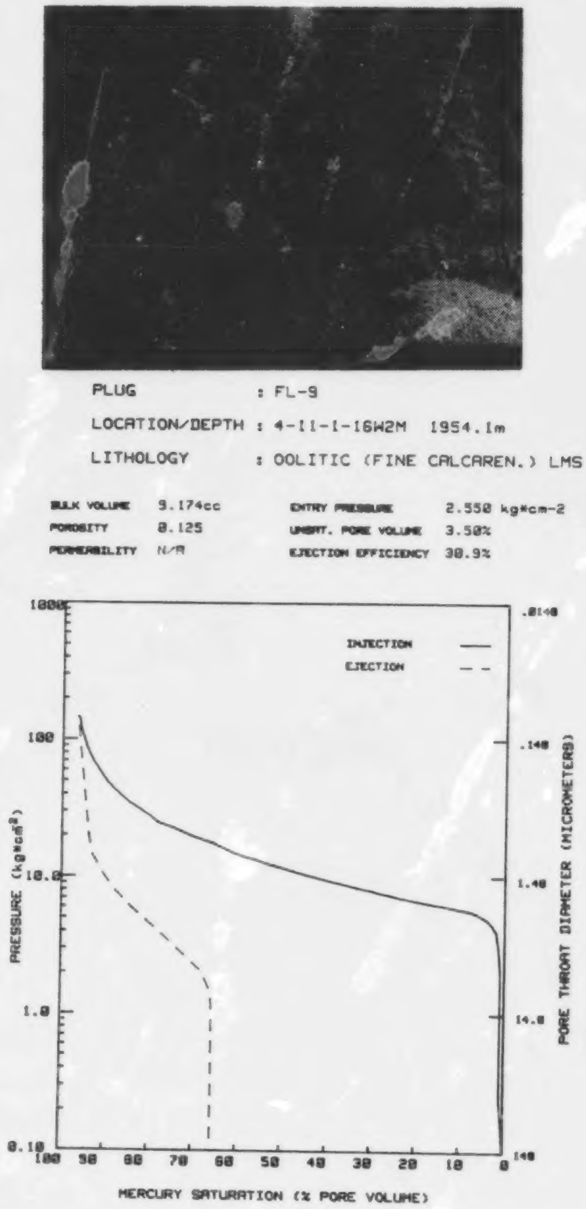
PLUG : FL-20

LOCATION/DEPTH : 4-4-1-16W2M 1977.1m

LITHOLOGY : MICRITIC DOLOMITE

BULK VOLUME	7.631cc	ENTRY PRESSURE	1.900 kg*cm-2
POROSITY	0.262	UNSAT. PORE VOLUME	13.20%
PERMEABILITY	20.5md	EJECTION EFFICIENCY	77.0%

Figure 14. Scanning electron micrograph and MICP profile of partially dolomitized rock with abundant micrite. Specimen FL-20 (4-4-1-16W2M; 1977.1 m). Scale bar is 20 micrometres.



PLUG : FL-9

LOCATION/DEPTH : 4-11-1-16W2M 1954.1m

LITHOLOGY : OOLITIC (FINE CALCAREN.) LMS

BULK VOLUME	9.174cc	ENTRY PRESSURE	2.550 kg*cm-2
POROSITY	0.125	UNSAT. PORE VOLUME	3.50%
PERMEABILITY	N/A	EJECTION EFFICIENCY	30.9%

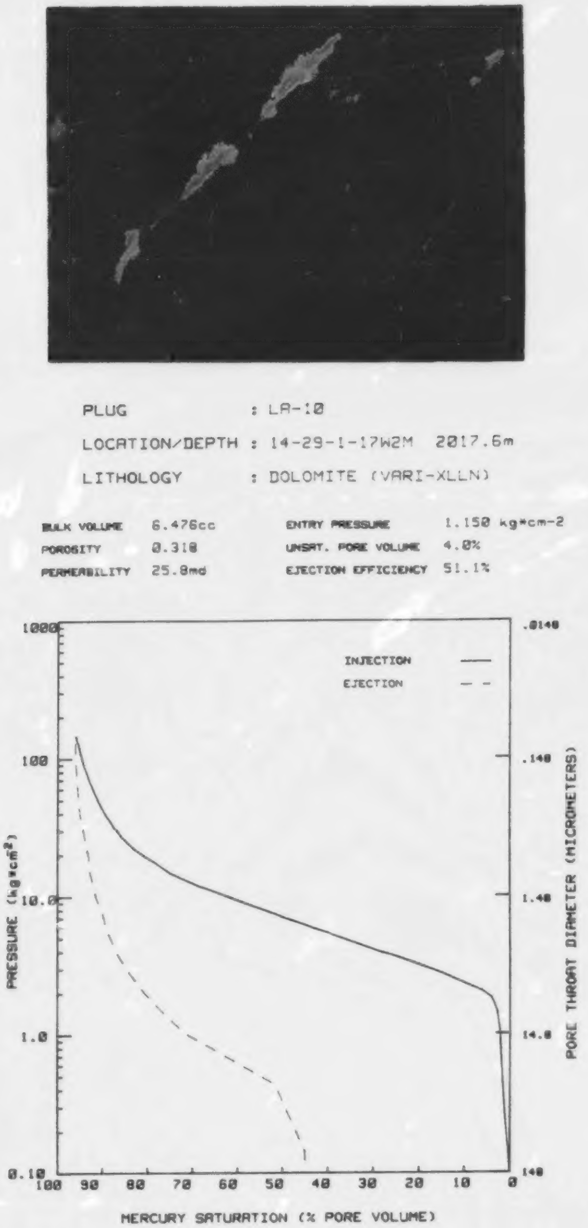
Figure 15. Scanning electron micrograph and MICP profile of Specimen FL-9 (4-11-1-16W2M; 1954.1 m). Oolitic wackestone with intergranular porosity. Scale bar is 100 micrometres.

LAKE ALMA FIELD (TOWNSHIP 1, RANGE 17W2M)

As at the Oungre Field, production at Lake Alma is from the Oungre zone carbonates. The trap at Lake Alma, however, is structural, with over 40 m (130 ft) of closure on the Ratcliffe Beds.

Reservoir lithologies at Lake Alma are microsucrosic dolomites, micrite-bearing dolomites, and rarely dolomitic lime wackestones to mudstones. Porous and permeable units are 1 to 2 m (3 to 7 ft) thick, interbedded with non-reservoir rock of similar or greater thickness.

Lake Alma reservoir rocks are characterized by high average porosity and permeability, a product of effective dolomitization, and hence ejection efficiencies are also high (Table 1). The addition of moldic and microvuggy porosity to the dolomite intercrystalline network has had no discernible effect on production capabilities. It may, however, contribute to high standard deviation of pore throat sizes, indicated by steeper slopes on MICP profiles (Fig. 16). Variability in size of dolomite crystallinity also adds to pore system heterogeneity.



PLUG : LA-10

LOCATION/DEPTH : 14-29-1-17W2M 2017.6m

LITHOLOGY : DOLOMITE (VARI-XLLN)

BULK VOLUME	6.476cc	ENTRY PRESSURE	1.150 kg*cm-2
POROSITY	0.318	UNSAT. PORE VOLUME	4.0%
PERMEABILITY	25.8md	EJECTION EFFICIENCY	51.1%

Figure 16. Scanning electron micrograph and MICP profile of Specimen LA-10 (14-29-1-17W2M: 2017.6 m). Micritic,variably crystalline dolomite with moldic and microvuggy porosity. Scale bar is 100 micrometres.

Non-reservoir intervals of the Oungre zone at Lake Alma Field are usually either calcite- or anhydrite-cemented oolitic-pelletoidal-skeletal packstones/grainstones, or impermeable, burrowed and argillaceous lime mudstones/wackestones. In contrast to the Oungre Field, where the packstone/grainstone lithology occurs as thin beds interpreted as storm deposits, at Lake Alma these sediments form discrete, thick units (2 to 5 m, 6 to 16 ft) interpreted as shoals. None of these rocks serve as reservoir and none are dolomitized, due to early occlusion of pore space by calcite or anhydrite. The abundance of 'grainy' rocks at Lake Alma indicate the existence of a higher energy depositional regime, at least periodically.

Moldic porosity, after skeletal debris, is much more common in the Lake Alma Field than in the Oungre Field, suggesting that more normal marine salinities prevailed in the Lake Alma area during Oungre zone sedimentation.

SUMMARY AND CONCLUSIONS

Dolomite and dolomitic limestone are the dominant reservoir rocks in the Ratcliffe Beds of southeastern Saskatchewan. The most common porosity type is intercrystalline. Intergranular, moldic and microvuggy porosity make only a minor contribution to the productivity of any given reservoir lithology.

The distribution of specific reservoir lithologies and associated pore systems is controlled by depositional factors which influence the degree of dolomitization. Variation in the character of the original lime sediments is related to position on the carbonate shelf and thus the energy level in a particular environment.

At the Oungre Field, the very fine crystallinity of the dolomite reservoir rocks causes them to be extremely sensitive to argillaceous material in the pore space. In fact, clay occlusion creates the up-dip permeability barrier.

At Flat Lake, where thick reservoir facies developed on structural highs, flanking very low permeability, 'chalky' sediments are largely responsible for constraining the pool to its present position. Though dolomite intercrystalline porosity is common, partial dolomitization of sediments with varying proportions of allochems plus moldic and vuggy porosity has introduced some heterogeneity into pore systems in Flat Lake reservoir.

Structure alone explains the oil accumulation at the Lake Alma Field, but subtle differences in the pore systems related to depositional facies are also recognizable.

ACKNOWLEDGEMENTS

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PERMIAN TO JURASSIC REDBEDS OF THE WILLISTON BASIN

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ABSTRACT

Redbeds of Permian to Jurassic age are present in outcrops in the Black Hills of South Dakota. Redbeds, which have been referred to as Jurassic in age, are present in outcrops in southern Manitoba. In the subsurface, equivalent strata are the Spearfish Formation of Triassic or Permo-Triassic age in North Dakota and the Lower Watrous or Amaranth Formation of Jura-Triassic or Jurassic age in Saskatchewan and Manitoba. Paleontological data are lacking for age resolution. A careful examination of physical evidence from the subsurface indicates that there are previously unrecognized regional unconformities within the clastic redbeds. Areal distribution patterns of these lithogenetic units favour the Jurassic age assignment for a part of the Lower Watrous and Amaranth Formations.

The Spearfish Formation, as originally defined, includes redbeds which have been correlated with strata from Permian to Jurassic age in eastern Wyoming, although it was originally generally assigned a Triassic age. The Spearfish Formation will retain its greatest utility in the subsurface if it is restricted to strata of Permian to Triassic age. Redefinition of the Belfield and Pine units would avoid the introduction of new names and by applying them to the two lithogenetic units within the Spearfish would more accurately portray the depositional history of these strata. Similarly a redefinition of the Saude, closely resembling the original definition, would preserve the utility of that term. The Saude, which bears an onlap relationship to underlying strata, would then be equivalent to the upper shaly unit of the lower Amaranth of Manitoba.

REGIONAL GEOLOGY AND PETROLEUM POTENTIAL OF THE LOWER AMARANTH FORMATION, COULTER - PIERSON AREA, SOUTHWESTERN MANITOBA

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ABSTRACT

The lower member of the Amaranth Formation in the Manitoba portion of the Williston Basin produces oil from sandstone subunits discontinuously developed in its lower part. In the Coulter-Pierson area current Lower Amaranth production represents about 7 percent of Manitoba's total oil production.

The Lower Amaranth is separated from underlying Mississippian strata by a marked angular unconformity, and is overlain by Jurassic Upper Amaranth evaporites. The member is subdivided into an upper shaly unit consisting mainly of mudstones and a lower sandy unit composed of siltstones and primarily fine- to medium-grained sandstones. The lower sandy unit comprises four sandstone subunits, designated in ascending order as 'Lower' Sand, 'Main' Sand, 'A' Sand and 'B' Sand.

The Lower Amaranth sediments were deposited in a low energy, tidal flat environment. Three generalized lithofacies are recognized: 1) supratidal mudstone facies; 2) intertidal siltstone facies; and 3) subtidal channel-fill sandstone facies.

The sandstone unit is the reservoir rock and the 'Main' Sand is the primary producer (2-5 m thick). Porosity is generally intergranular and variable due to anhydrite occlusion. Entrapment is primarily stratigraphic. Oil is trapped within the porous sandstones by porosity - permeability pinch-out resulting from lateral and vertical variations in lithofacies. Reservoir characteristics are heterogeneous and are controlled by Mississippian topography, depositional environment, compaction and anhydritization through diagenetic processes.

INTRODUCTION

The Lower Amaranth member of the Amaranth Formation is present over a large area of the Manitoba portion of the Williston Basin (Figs. 1, 2). Monthly oil production in excess of 17,000 m^3 (107,000 bbl) is obtained from discontinuous, fine grained sandstone and siltstone beds in the lower portion of the member in the Waskada and South Pierson Fields in southwestern Manitoba. Limited production also occurs in and around the Coulter Field. This oil production accounts for approximately 27 percent of the monthly (March, 1990 data) production of 62,840 m^3 (395,260 bbl) in southwest Manitoba. The remaining 73 percent is produced from Mississippian rocks.

STUDY AREA

The study area is located in the extreme southwest corner of Manitoba on the northeast flank of the Williston Basin (Fig. 1). It includes 9 townships, (Twp. 1-3, Rge. 27-29; Fig. 2) and is bounded by the international border to the south and the Manitoba-Saskatchewan border to the west. In this report the study area is also referred to as "Coulter-Pierson" area.

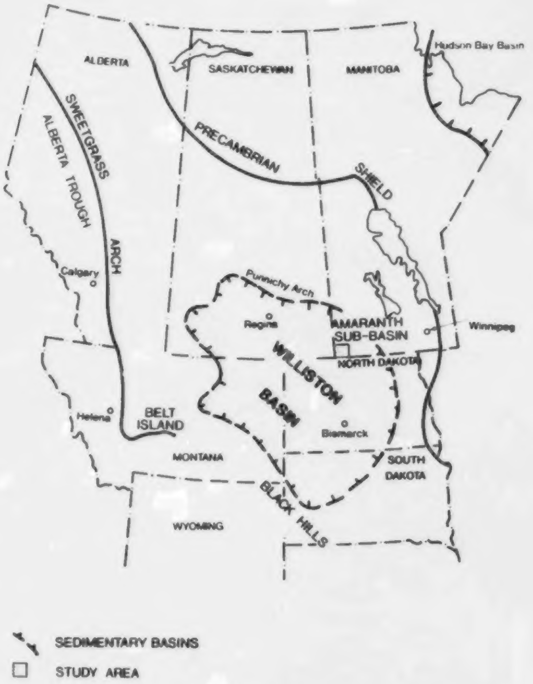

Figure 1: Location of the Williston Basin and major tectonic elements. (Modified from: Carlson, 1968; Christopher, 1984).

EXPLORATION AND PRODUCTION HISTORY

Oil production from the Mississippian Mission Canyon Formation has been continuous for 30 years in southwestern Manitoba. Exploration for oil in the Lower Amaranth of the Coulter-Pierson area commenced as a result of discovery and subsequent development of the Waskada Lower Amaranth A Pool in 1980 by Omega Hydrocarbons Ltd. The geology and petroleum potential of the Waskada-Pierson area was described by Barchyn (1982).

Three oil fields, Coulter, Pierson and South Pierson, have been designated in the study area. Lower Amaranth oil production is restricted mainly to the Coulter and South Pierson fields. Production in the Pierson Field is from the Mission Canyon Formation.

In September of 1981 Lyleton Corporation discovered oil in the Lower Amaranth member at the 12-30-2-28 WPM well and established the first Lower Amaranth Pool in the South Pierson Field. Previous production from the field was from the Mississippian Mission Canyon Formation. Because of low productivity and completion difficulties, development of this pool was limited.

In December, 1982, Newscope Resources Ltd. drilled and completed the 11-21-1-27 WPM well, thereby establishing the first

Lower Amaranth Pool in the Coulter Field. In the first 12 month period the well produced 238 m^3 of oil and 4,261 m^3 of water (94 percent water cut). Subsequent to this discovery seven wells were drilled in the pool within a one year period. Average water cut in the Coulter Field is 88 percent.

In 1985, Home Oil Company made a significant discovery in the Lower Amaranth at 16-8-2-29 WPM well, with initial production rate of 9.9 m^3/d (62 bopd), a very high rate for Manitoba production. This resulted in an active delineation and development program in the South Pierson Field. To date (June 30, 1989) Home has drilled 23 wells in the area; 16 are producing.

The exploration and development history of each pool in the study area is described in Table 1.

Table 1: Exploration and Development History

FIELD	POOL	POOL CODE	DISCOVERY DATE	DISCOVERY WELL	NO. OF WELLS DRILLED	PROSPECTS FOR ADDITIONAL DEVELOPMENT*
Coulter	A	11 29A	Dec/82	11-21-1-27	8	6
South	A	12 29A	Sept/81	12-30-2-28	7	4
Pierson	B	12 29B	Dec/85	16-08-2-29	14	19
	C	12 29C	June/87	06-19-2-29	2	3
	D	12 29D	Jan/88	04-01-2-29	1	2
	E	12 29E	Dec/86	06-07-2-29	1	2

* based on 16 hectare spacing

TERMINOLOGY

The term "Amaranth" was first proposed by Kirk (1929) for the reddish-brown siltstones and dolomitic shales that occur near the town of Amaranth, Manitoba (Fig. 2). This sequence occurs at the base of the Mesozoic section. It was later traced by Wickenden (1945) from the type locality into the subsurface and was formally defined as the "Amaranth Formation" at 2-9-8-26 WPM and 14-15-9-3 WPM wells. The Amaranth sequence was subdivided by Stott (1955) into the Upper Amaranth Unit, composed mainly of evaporites, and the Lower Amaranth Unit, comprising mudstone in its upper part and siltstone with sandstone in its lower part. McCabe (1956) redefined the "unit" as a "member" and Hansen (1987) followed the "member" terminology. Barchyn (1982), Rodgers (1986), and Husain and Halabura (1987) described the Lower Amaranth sequence not as a "lower member" of the Amaranth Formation, but rather as the "Lower Amaranth Formation".

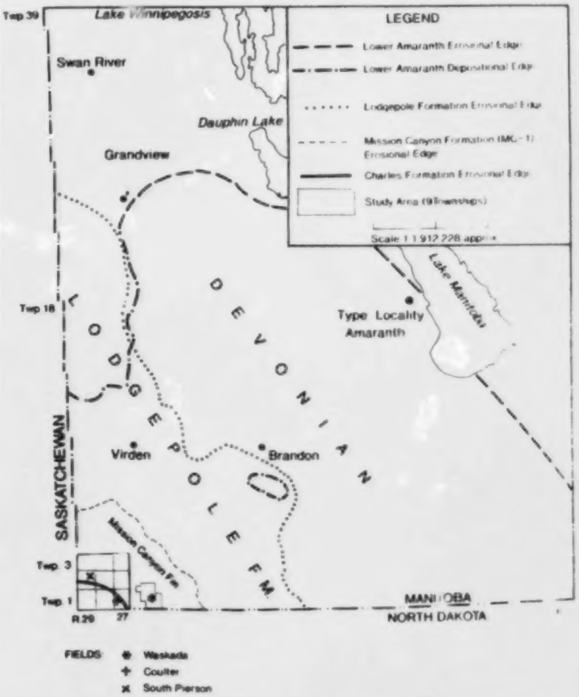

Figure 2: Location of study area and distribution (subcrop edges) of Amaranth and Paleozoic formations in southwestern Manitoba. (Modified from: Stott, 1955; McCabe, 1956; Hansen, 1987).

REGIONAL GEOLOGICAL SETTING

The Williston Basin extends over southeastern and south-central Saskatchewan, the southwestern corner of Manitoba, most of North Dakota, northwestern South Dakota and northeastern Montana (Fig. 1). The sedimentary strata comprise three major lithological successions separated by major angular unconformities. The lithological sequences are: a) Paleozoic clastics, carbonates and evaporites, b) Mesozoic clastics carbonates and evaporites, and c) Cenozoic clastics. Basinward thickening of sediments contemporaneous with basin subsidence has resulted in the formation of a sedimentary wedge that in Manitoba dips to the southwest (McCabe, 1959).

During the Jurassic, the Williston Basin was separated from the Alberta trough by the Sweetgrass Arch, bordered to the north by the Punnichy Arch in southern Saskatchewan and to the east by the Precambrian Shield (Christopher, 1984; Fig. 1). The Williston Basin attained several different configurations from late Paleozoic to early Cretaceous (Hansen, 1987). During deposition of the Amaranth Formation the basin had two distinct elements. A northeasterly-trending trough or sub-basin (extending from northern Wyoming and South Dakota to North Dakota and into southwestern Manitoba) formed the Amaranth Embayment or sub-basin (Fig. 1). A second trough extended northwest from the depositional centre of the basin (in North Dakota and northern Montana) into southern Saskatchewan (Christopher, 1984; Hansen, 1987). In pre-Lower Amaranth time, the Manitoba portion of the Williston Basin was a very low-lying, uniform, peneplained surface with few minor ridges and valleys controlled primarily by the structure of the underlying strata (McCabe, 1956).

A structure map (Fig. 3) displays relief on the Mississippian erosional surface subjacent to the Lower Amaranth member. This unconformable surface dips gently basinward (to the southwest). Thickness of the Lower Amaranth is (Fig. 4) controlled by the Mississippian topography and possibly also by minor syndepositional tectonic activity (McCabe, 1956).

STRATIGRAPHY

The lower member of the Amaranth Formation is the lowermost part of the Jura-Triassic succession in Manitoba. Based on regional correlation and detailed lithological comparison the Lower Amaranth member is considered to be stratigraphically equivalent to the Lower Watrous of southeast Saskatchewan and the Spearfish Formation of North Dakota. As such, these formations are considered the oldest rocks of the Mesozoic sequence in the Williston Basin (Francis, 1957; Christopher, 1984). Because of the lack of diagnostic fossils, the age of the Lower Amaranth, and its stratigraphically equivalent strata, has

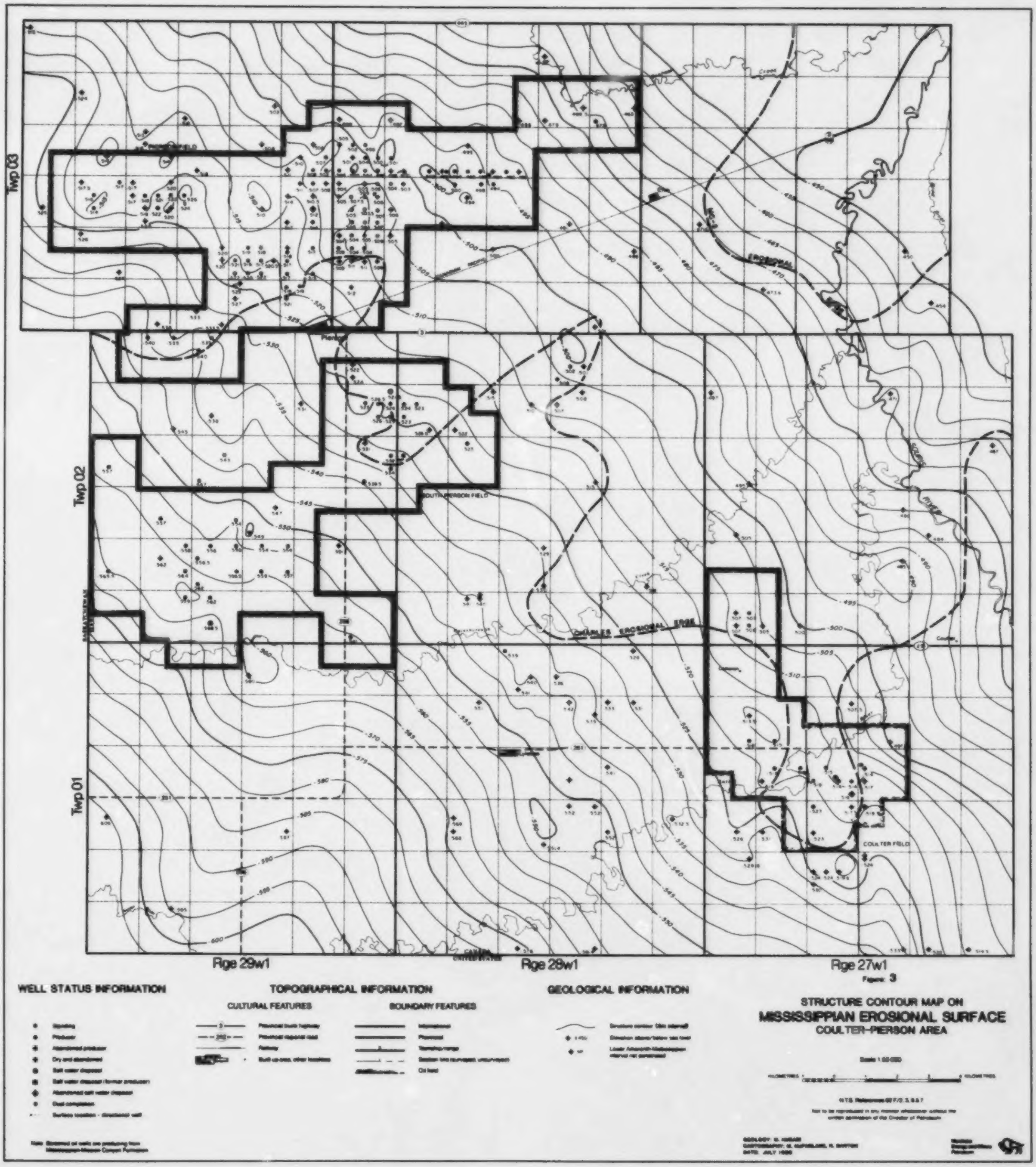

Figure 3: Structure contour map of Mississippian erosional surface.

been a subject of continuous debate for fifty years (Wickenden, 1945; Imlay, 1947; Schmitt, 1953; McCabe 1956; Francis, 1957; Carlson, 1968; Christopher, 1984). More recently Lower Amaranth, Lower Watrous and Spearfish strata have been placed in the Triassic of the Williston Basin Stratigraphic Nomenclature Chart (Geological Society of America, 1987).

In southwestern Manitoba the Lower Amaranth unconformably overlies Ordovician, Silurian, Devonian and Mississippian (Fig. 2) rocks. In the southern and southwestern portion of the study area the underlying bed is a thin erosional wedge of Mississippian Charles strata composed of anhydrite and anhydritic dolomite, whereas in the north and northeastern portion it is the MC-3 and members of the Mississippian Mission Canyon Formation (Figs. 2, 3, 4).

The topography of the Mississippian erosional surface controls the distribution and thickness variation of reservoir rocks in the Lower Amaranth. This thickness varies from a minimum of 25 m in the centre of the Pierson Field to a maximum of 46 m in the south of the Coulter Field. Thickness generally increases basinward. Several elongated, sub-parallel belts indicated in the isopach map show an overall northeast trend (Fig. 4).

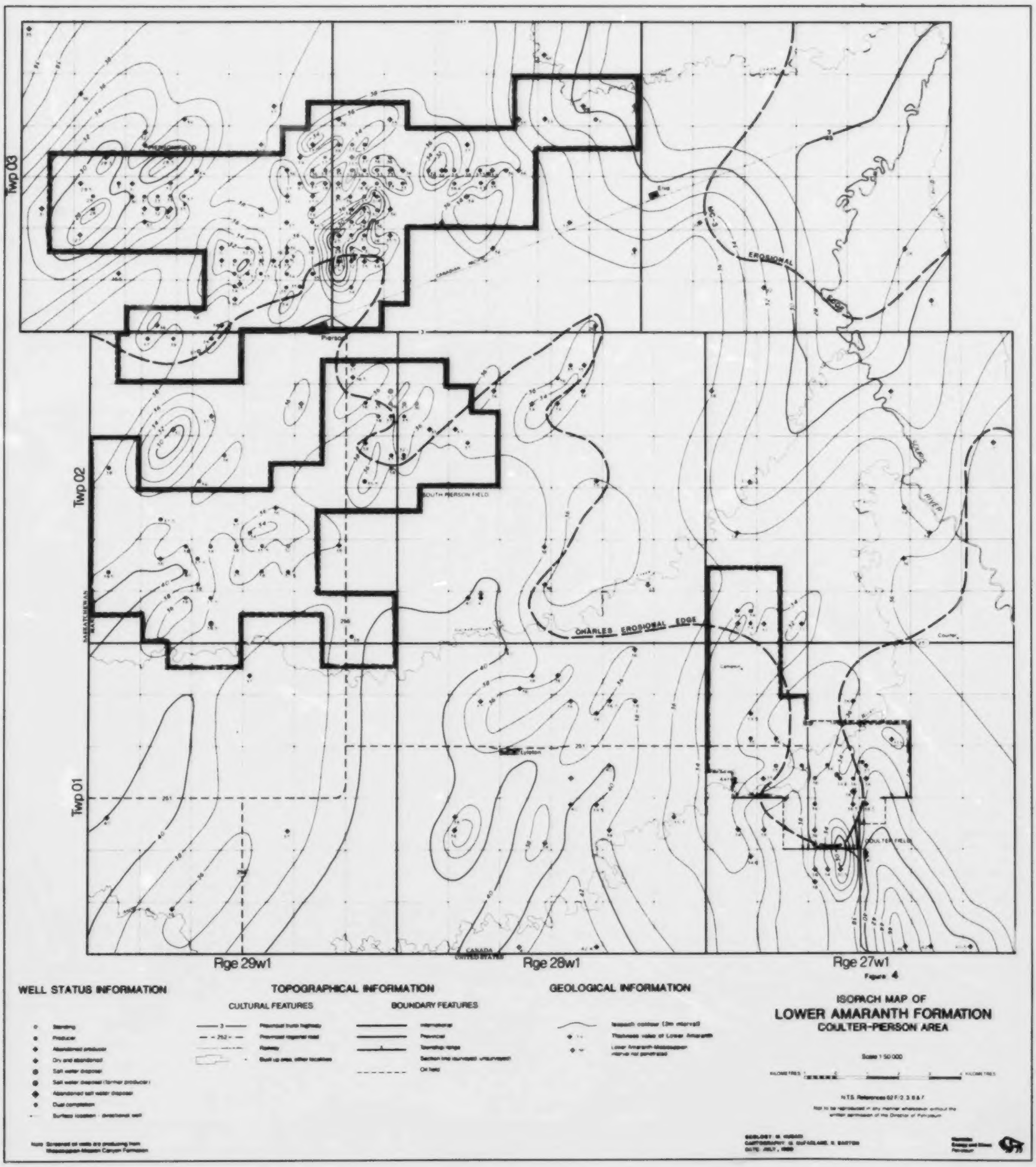
Twp 03
Twp 02
Twp 01
Rge 29w1
Rge 28w1
Rge 27w1
Pierson
SOUTH PIERSON FIELD
CHARLES EROSIONAL EDGE
EROSIONAL EDGE
MC-3
RIVER
Lyleton
COULTER FIELD
Coulter
CANADA
UNITED STATES
Figure 4
WELL STATUS INFORMATION
Standing
Producer
Abandoned producer
Dry and abandoned
Salt water disposal
Salt water disposal (former producer)
Abandoned salt water disposal
Dual completion
Surface location - directional well
Note: Screened oil wells are producing from Mississippian Mission Canyon Formation
TOPOGRAPHICAL INFORMATION
CULTURAL FEATURES
Provincial trunk highway
Provincial regional road
Railway
Built up area, other locations
BOUNDARY FEATURES
International
Provincial
Township range
Section line (surveyed, unsurveyed)
Oil field
GEOLOGICAL INFORMATION
Isopach contour (2m interval)
Thickness value of Lower Amaranth
Lower Amaranth-Mississippian interval not penetrated
ISOPACH MAP OF
LOWER AMARANTH FORMATION
COULTER-PIERSON AREA
Scale 1:50 000
KILOMETRES
N.T.S. References 62 F/2, 3, 6 & 7
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Figure 4: Isopach map of Lower Amaranth member.

LITHOLOGY AND CHARACTERISTICS

In the study area the lower member of the Amaranth Formation comprises mudstone, siltstone and sandstone sequences. The formation is subdivided into two distinct units, an upper shaly and a lower sandy unit (Fig. 5).

The main characteristics of the formation as noted in core from the study area are: 1) reddish-brown colour, 2) wavy- to lenticular-bedding, 3) anhydrite nodules and blebs, 4) scattered siltstone lenses and/or interclasts, 5) parallel- and/or cross-laminations and 6) locally preserved mud cracks.

LOWER SANDY UNIT

The lower sandy unit ranges in thickness between 6 and 20 m. In township 3 of range 27-29 WPM it is from 6 to 12 m thick, whereas in the southern part of the map area it is approximately 12 to 20 m thick.

The lower sandy unit consists of reddish-brown siltstone and sandstone with anhydrite blebs and nodules and silty mudstone stringers. The sandstone is generally very fine- to fine-grained and locally medium- to coarse-grained. The siltstone is commonly massive with fine grained sandstone stringers and/or intraclasts. Mudstone stringers are also common in the upper part of the unit.

The main characteristics of the unit are: 1) gross lithology siltstone, 2) lenticular bedding, 3) abundance of anhydrite nodules and blebs, 4) thin shale laminae, 5) up to four sandstone subunits in a fining upward sequence separated by siltstone beds (Figs. 5, 6).

In the study area the lower sandy unit contains up to four sandstone subunits of variable thickness (0.5 to 5.0 m) designated "lower", "main", "A" and "B" sands (Fig. 5). Each subunit is separated by tight siltstone beds. Where the "B" sand is present its top defines the upper contact of the lower sandy unit; where it is not developed the top of the unit is picked from the siltstone response in mechanical logs. Individual sandstone subunits are illustrated in the log for the 4-1-2-29 well, which exhibits the best sandstone development in the study area (Fig. 5).

Based on regional mapping and stratigraphic cross-sections the "lower", "A" and "B" sandstone subunits are discontinuously distributed in the study area. These sandstone bodies are not generally stratigraphically equivalent and therefore not necessarily correlatable. Each of the sandstone subunits grades laterally to siltstone or interbedded shale and sandstone (Barchyn, 1982).

The "main" sandstone subunit is generally present throughout the study area, and is composed of very fine- to fine-grained sandstone to siltstone to shaly siltstone. The quality (siltiness and shaliness) of this subunit deteriorates, and its thickness thins to the north in township 3 and in areas outside the established pools.

UPPER SHALY UNIT

In the study area the thickness of the upper shaly unit ranges from a minimum of 13 m to a maximum of 26 m. The unit is generally composed of reddish-brown to greenish-brown mudstone. Bands of dolomitic and shaly siltstone occur close to the contact with the lower sandy unit. Based generally on gamma ray log response the percentage of siltstone appears to be variable and in some wells the upper unit is devoid of siltstone.

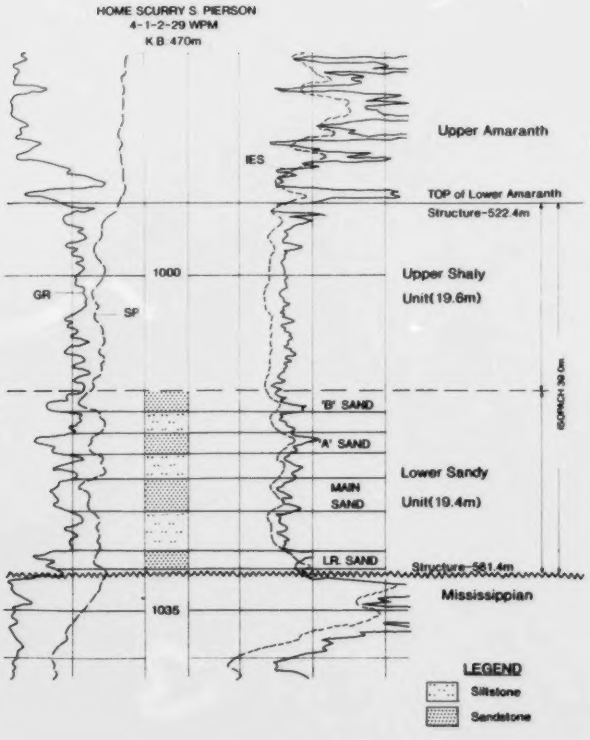

Figure 5: Type log of Lower Amaranth member showing subdivisions and best sand development in the Coulter-Pierson area. (Modified from: Barchyn, 1982).

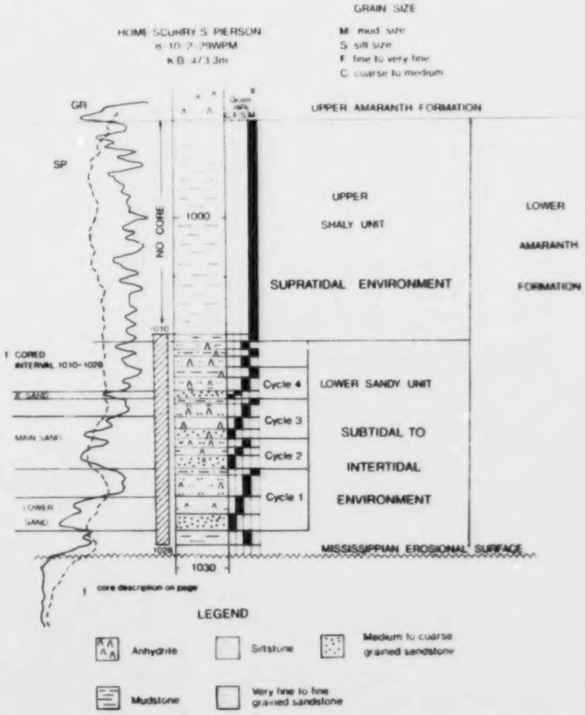

Figure 6: Depositional environment and rock sequence of Lower Amaranth member showing repeated cycles of fining upward sequences in the Coulter-Pierson area. (Modified from: Klein, 1977).

The main characteristics of the unit are: 1) gross lithology mudstone, 2) dolomitic and shaly siltstone stringers common in lower part, 3) wavy bedding at places, 4) scattered anhydrite blebs in lower part, 5) parallel laminations and 6) scattered siltstone lenses and/or intraclasts in lower part.

RELATIONSHIPS BETWEEN UPPER SHALY AND LOWER SANDY UNITS

1. colour of both units is the same;
2. the contact between the two units is transitional;
3. percentage of silt, sand and anhydrite decreases from lower sandy unit to upper shaly unit;
4. lower sandy unit has lenticular bedding whereas the lower portion of the upper shaly unit has wavy bedding; and
5. Hansen (1987) reported local mud cracks in the upper shaly unit.

STRUCTURE

In southwestern Manitoba a major erosional unconformity lies between Mesozoic and Paleozoic strata. Structure maps on the geological formations are regionally similar. However, the effects of local tectonic activity and/or solution of the Devonian Prairie (salt) Formation and subsequent collapse of overlying formations can be deduced from the isopachs of these formations (McCabe, 1956, 1959). Structural highs and lows in the Mesozoic sediments could be due to differential compaction and draping over local topographic highs on the erosional surface. Mesozoic strata strike northwest with a regional dip to the southwest at approximately 7 m/km.

Topographic contours on the Mississippian surface reflect a combination of paleotopographic relief on the unconformity, and later tilting and possible deformation (Fig. 3). Those in Sections 7 and 18 of Township 3, Range 28 WPM show a closed high in association with the Mission Canyon oil pools in the Pierson Field and a few isolated one-well closures are found in the map area. These anomalies are not reflected by structure on the top of the Lower Amaranth indicating them to be related to only Mississippian topography.

The paleotopography of the Mississippian erosional surface in the study area is best illustrated by the isopach of the overlying Lower Amaranth (Fig. 4). The top of the Lower Amaranth approximates a time-stratigraphic marker (McCabe, 1956, 1959), and consequently topographic highs on the Mississippian surface coincide with the thinner parts of the Lower Amaranth (Sections 7, 18; Township 3, Range 28 WPM). However, caution must be used in interpreting the isopach anomalies as paleotopographic features, because other factors such as local uplift or subsidence during Lower Amaranth time, and differential compaction and/or salt solution and subsequent collapse of the underlying Mississippian formations contemporaneous with Lower Amaranth deposition can also affect Lower Amaranth thicknesses. It is noted that sand thickness is directly related to the thickness of the Lower Amaranth member.

DEPOSITIONAL ENVIRONMENT

In southwestern Manitoba the need for a depositional model is acute because oil accumulations in the Lower Amaranth are likely controlled by subtle stratigraphic factors rather than by structure.

PREVIOUS INTERPRETATIONS

Stott (1955) interpreted the Lower Amaranth rock sequence as a restricted shallow marine deposit. Arid conditions during Lower Amaranth time resulted in preservation of residual iron compounds that gave the rock its red colour. Normal circulation of sea water was hindered by numerous remnants of old limestone beds, and under restricted conditions calcium sulphate was precipitated around the perimeter of the Manitoba portion of the Williston Basin. Absence of coarse clastics indicated that the surrounding land mass was low lying. The red beds were derived from reworking of older red beds (Lyleton and Ashern formations).

McCabe (1956) suggested an arid climate under a terrestrial oxidizing environment rather than a marine environment for deposition of the Lower Amaranth. He attributed the presence of contorted structures with irregular patches of fine bedded and cross-bedded sediments, and slickenslides to be probably indicative of a terrestrial or transitional environment. The pronounced frosting and pitting of most of the sand grains support a terrestrial source. The roundness, sphericity and well sorted nature of the sand grains and presence of some unaltered feldspar grains are indicative of arid aeolian transport. He believed anhydrite in the form of irregular to rounded patches to be clastic and apparently derived from break-up of primary anhydrite layers somewhere on the land mass.

Barchyn (1982) invoked an environment ranging from dominantly continental to marine under intermittent shallow restricted conditions representing the initial stages of a major transgression of Jurassic seas. He suggested that the cyclical fining upward sequence of sandstone beds were fluvial deposits on an intermittently emergent mud flat. The occurrence of anhydrite nodules suggests that hypersaline conditions were closely associated with deposition or else they were products of diagenesis. An alternative depositional mechanism could have been winnowing of sediments by wave action in shallow marine conditions.

Hansen (1987) compared the depositional environment of the Lower Amaranth member with that of a modern low-energy tidal flat on the Colorado Delta. Here, silty clay and fine grained sands are supplied to the north coast of the Gulf of California by the Colorado River. A featureless mud plain grading into a broad, shallow, low-dipping, subtidal plain has developed within a prevailing arid climate. Sediments deposited in the subtidal to lower intertidal zone are composed of laminated silty clay with thin (1 mm) bands of silt to fine sand at regular intervals. The intertidal zone is dominated by wavy, interlaminated silty clay, silt and shaly silt.

Accordingly, the Lower Amaranth rocks were deposited on a broad featureless tidal flat, similar to the setting along the northeast coast of the Gulf of California, under the following conditions:

1. The basal reddish-brown siltstones overlying the Mississippian unconformity were deposited in a supratidal mud flat to sabkha type setting along the restricted margin of the Manitoba portion of the Williston Basin. Nodular anhydrite was formed by displacive growth in the sediments.
2. Very fine grained subarkose was deposited in very shallow water below wave base.
3. Locally developed fining upward coarse to medium grained subarkose was deposited in tidal channels.
4. Gradual recession of water from the tidal flat resulted in deposition of the wavy to lenticular bedded siltstones with very fine grained sandstones and mudstones associated with

fluctuating current energy in an intertidal zone. Wavy to lenticular bedding structures formed under the influence of small waves and tidal currents on the tidal flat.

5. Aeolian processes responsible for 'frosted' quartz grains remained active throughout deposition of the lower sandy unit.
6. With further progradation of the tidal flat, mudstones with minor siltstones of the upper shaly unit were deposited in a supratidal mud flat setting. This setting was, in part, subaerially exposed and locally mudcracked.
7. Sedimentation of the upper shaly unit was eventually terminated. Seas became increasingly hypersaline and the extensive anhydrite sequence of the Upper Amaranth member was precipitated.

Kooyman (1989) suggested a tidal flat depositional environment for rock sequences of the Lower Amaranth in southern Manitoba.

CURRENT INTERPRETATIONS

Three basic lithofacies have been recognized in the Lower Amaranth in the study area:

a) sandstone facies - locally developed within the siltstone facies - lower sandy unit,
b) siltstone facies with minor mudstone - lower sandy unit, and
c) mudstone facies with minor siltstone - upper shaly unit

Sandstone Facies: (Subtidal Deposits)

The basic sedimentary characteristics of this lithofacies are: 1) up to four sandstone subunits in a fining upward sequence, 2) sandstone grains mostly subrounded to rounded and rarely subangular and range in size generally from fine to very fine and locally from coarse to medium, 3) crossbedding, 4) thickness variations, 5) sandstones discontinuous (and lens shaped), occurring at different stratigraphic positions, 6) abundance of subrounded to rounded noncemented 'frosted' medium to coarse quartz grains, 7) abundance of anhydrite nodules and/or blebs, and 8) local occurrence of siltstone intraclasts and/or lenses.

In this facies fining upward sequences are noted in the drill core and are repeated to a maximum of four times in four individual cycles as observed in 4-1-2-29 WPM and 8-10-2-29 WPM wells (Figs. 5, 6). The above characteristics are similar to those noted by Hansen (1987) and could be the products of deposition in tidal channels in the subtidal environment.

Progradation of tidal flats tends to produce a fining upward sequence that reflects a transition from low tide sand flats, to high tide mud flats, and eventually to supratidal flats. This sequence may be cut at any level by erosive tidal channel sequences (Reineck, 1967). The abrupt presence or absence of the tidal-channel coarse grained sandstone facies of the lower sandy unit could be explained by this process.

Siltstone Facies: (Intertidal Deposits)

This facies consists of brownish-red, interlaminated siltstones with very thin beds of mudstones in its upper part, and very fine- to fine-grained sandstones in the lower part. Lenticular bedding is common. Local siltstone and mudstone intraclasts are also noted. Thin shale laminae occur in the upper part of the unit. Frosted, noncemented, medium to coarse quartz grains are common in the lower part. Anhydrite nodules and blebs occur throughout.

The characteristics described above are similar to those noted by Hansen (1987) and could suggest an intertidal environment for this facies with several cyclic (intermittent) breaks that resulted in local channel-fill deposits of the sandstone facies in the form of individual subunits. A transitional sequence of this facies, depending upon its sand/silt ratio, can be deposited anywhere from upper subtidal to lower intertidal flats (Hansen, 1987).

Mudstone Facies: (Supratidal Deposits)

The upper shaly unit is a mud dominated sequence. This facies comprises reddish-brown mudstones and shales with local bands of shaly siltstones and very fine grained silty, shaly sandstones in the lower portion of the sequence. The upper part is homogeneous, massive and thick bedded compared to the lower part where parallel-laminations and wavy bedding are noted. Scattered anhydrite blebs and siltstone lenses and intraclasts occur in the lower portion of the sequence. Local mud cracks are also reported in this facies (Hansen, 1987).

These sedimentary characteristics are similar to those reported by Hansen (1987) and indicate that this facies is the product of deposition under weak, low energy, very shallow water conditions common in the supratidal zone.

At places the contact between the siltstone facies of the lower sandy unit and the mudstone facies of the upper shaly unit is gradational. Rocks that are composed of silt and mud in highly variable percentages become a transitional sequence which has wavy- to lenticular-bedding. This transitional sequence suggests depositional conditions that could range from upper intertidal to lower supratidal before supratidal prevailed.

In summary, the "lower sandy unit" was apparently deposited under conditions ranging from subtidal (channel-fill sandstone facies) to intertidal flats (siltstone facies), to supratidal mud flat "upper shaly unit".

Therefore it is suggested that the Lower Amaranth in the Coulter-Pierson area was deposited in a tidal flat environment (Fig. 7).

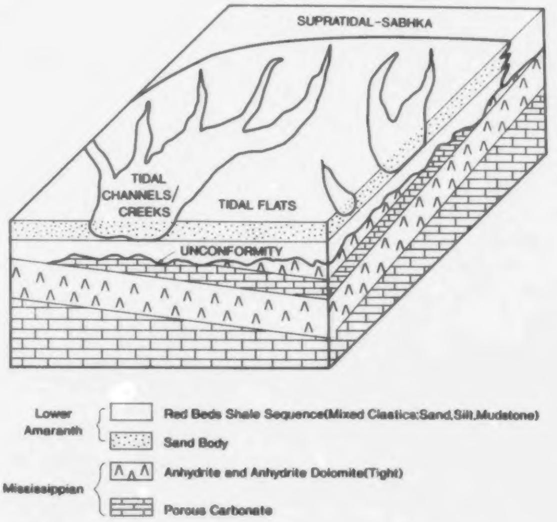

Figure 7: Depositional model of Lower Amaranth member in the Coulter-Pierson area. (Modified from: Weimer, 1981; Kooyman, 1989).

HYDROCARBON POTENTIAL

TRAPPING MECHANISM

Oil production in Manitoba is confined to the northeastern flank of the Williston Basin mainly as a result of the post-Mississippian pre-Mesozoic unconformity. Mississippian beds currently produce about 73 percent of Manitoba oil. The remaining 23 percent is obtained from discontinuously developed sandstone facies in the lower portion of the Lower Amaranth in the Waskada and Coulter-Pierson areas.

Paleotopography of the Mississippian erosional surface controls the distribution and thickness variation of the Lower Amaranth member in the study area (Figs. 3, 4). Thickness of the formation is positively correlated with the thickness of the lower sandy unit within which the reservoir facies (sandstone) is developed. It is therefore postulated that the paleotopography of the Mississippian erosional surface controls the distribution of the reservoir facies within which oil accumulations occur, depending upon development of favourable reservoir characteristics and parameters.

In the study area the trapping mechanism for Lower Amaranth oil is mainly stratigraphic with no known evidence for any structural trap (control). The reservoir is primarily controlled by facies distribution within the lower sandy unit, *i.e.*, porous sandstone pinching out against tight siltstone.

Underlying the Lower Amaranth sediments are Mississippian carbonates truncated by the post-Mississippian pre-Mesozoic (pre-Lower Amaranth) unconformity. The uppermost part of the Mississippian sequence is commonly characterized by dense anhydrite and anhydritic dolomite, which is identified as an "altered zone", extremely variable in thickness. It is generally assumed that the altered zone is formed by anhydritization and dolomitization of porous limestone through diagenesis (McCabe, 1959; Barchyn, 1982; Hansen, 1987).

Where it is present, the altered zone forms an effective seal for entrapment of hydrocarbon in the Mississippian carbonates. It is postulated that the oil trapped in the Lower Amaranth member is Mississippian oil that migrated upward into the Lower Amaranth reservoir facies in areas where the effective seal (altered zone) is lacking (thin or absent) and/or where the Lower Amaranth rocks, immediately overlying the unconformity, are permeable due to possible fractures (Barchyn, 1982; Hansen, 1987). Oil is thus accumulated within the porous sandstone facies and is trapped by the overlying impermeable siltstone/mudstone facies and laterally by the pinchout of porosity and permeability within the sequence of the lower sandy unit (Figs. 5, 6). The impermeable siltstones and mudstones prevent further vertical and updip migration. Consequently, the lower sandy unit acts as both a reservoir and a seal in most cases.

RESERVOIR FACIES

The main reservoir rocks of the Lower Amaranth in the study area are generally fine and locally medium- to coarse-grained quartzose sandstone present in the form of several individual subunits separated by impermeable siltstones in the lower sandy unit (Figs. 5, 6).

It must be noted that the subunits developed within each cycle, are not always composed of pure sandstone (generally quartzose). Rather the facies change from sandstone within the pools, to siltstone and/or silty mudstone outside the pool boundaries, in accordance with the prevailing environment that existed in the area at the time of deposition.

Figure 8 is an isopach map of the composite net pay of the potential reservoir sandstone present in each established pool. In constructing this map cutoffs of eight percent porosity and one millidarcy permeability (from core analysis) have been assumed. In general this map outlines the pool boundaries and possible extension of the oil-bearing reservoir.

RESERVOIR CHARACTERISITCS

The reservoir sandstone facies is variable in nature. Porosity is generally erratic resulting in differing reservoir characteristics in each designated pool. In the study area reservoir beds are heterogeneous and complex in terms of reservoir parameters. Arbez (1990) provides a detailed method for obtaining effective and total porosities and water saturations of the Lower Amaranth reservoir beds. Though Arbez's method is based on data from the South Pierson Lower Amaranth pools, the same approach may be applied elsewhere in the area to roughly estimate reservoir parameters.

POROSITY IN THE STUDY AREA

Porosity in the sandstones and siltstones is generally intergranular modified by authigenic minerals (anhydrites in most cases). Abundant anhydrites occur throughout the sequence including the reservoir sandstone facies. Secondary anhydrites have occluded the pre-existing porosity. Some of the cleanest sandstones (under the microscope and by gamma ray log response) show lower permeabilities due to anhydrite infilling pore space. Sandstones with nonspherical and subrounded to subangular grains tend to have higher visible porosities than sandstones composed of spherical and well rounded grains. Porosity varies erratically from 7 to 25 percent (core analysis) from well to well. With variation in porosity comes considerable variation in permeability.

Coulter Field

The reservoir rock in the Coulter Field is characterized by a silty, very fine to fine grained, and rare medium to coarse grained, quartzose sandstone facies with porosities ranging from 7 to 19 percent and permeabilities from 0.5 to 4.0 millidarcies. The reservoir grains are poorly sorted and authigenic anhydrite has occluded the intergranular pores thus reducing the porosities and permeabilities. Interlaminated siltstone lenses are also noted. Consequently, the reservoir porosity is unevenly distributed. The net pay in this field ranges from 2.0 to 3.5 m (Fig. 8).

South Pierson Field

The reservoir rocks in the South Pierson Field are generally composed of well-sorted, fine- to medium- to coarse-grained sandstones with clay and anhydrite free matrix resulting in the development of better porosities and permeabilities than those present in the Coulter Field. The textural variation is attributed to a localized high energy environment during channel-fill and channel-creek conditions. In the A Pool, the porosities are comparatively erratic, irregular and scattered, and permeabilities are comparatively low. In the B, C, D and E Pools good to excellent porosities are noted. Porosities range from 12 to 25 percent and permeabilities from 2 to 19 millidarcies. The net pay ranges from 2.5 to 4.5 m in the "A" pool and from 3.0 to 7.0 m in the "B" pool (Fig. 3).

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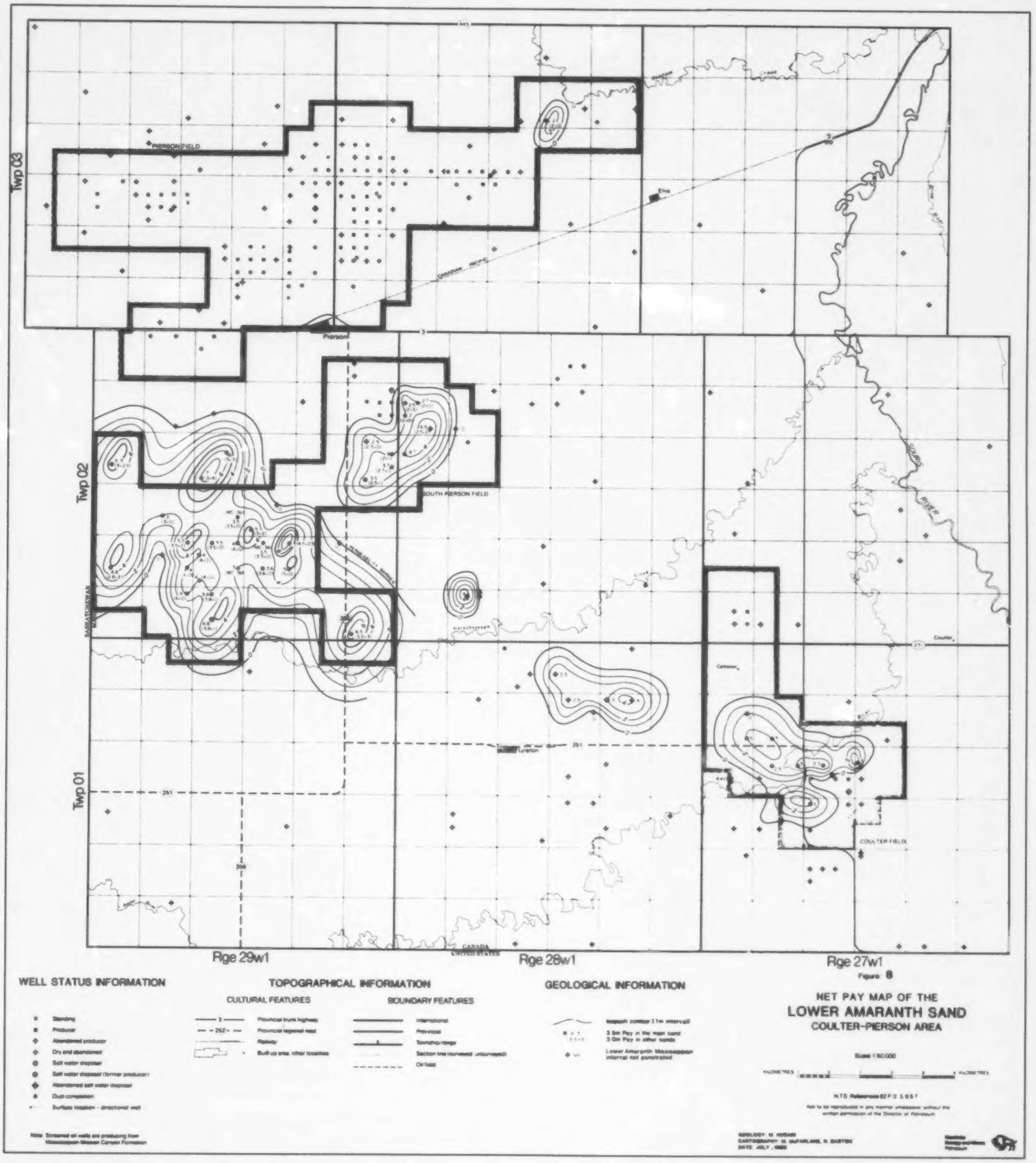

Figure 8: Isopach map of the net pay present in each Lower Amaranth pool of the Coulter-Pierson area.

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SEDIMENTOLOGY AND PETROLEUM GEOLOGY OF A TRIASSIC SPEARFISH SANDSTONE RESERVOIR, SOUTH STARBUCK FIELD, BOTTINEAU COUNTY, NORTH DAKOTA

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ABSTRACT

The South Starbuck field (Township 160N, Range 78W) produces from a Triassic Spearfish Formation point bar sandstone that rests unconformably upon evaporites and carbonates of the Mississippian Charles Formation or is in erosional contact with Spearfish bioturbated mudstones/siltstones. The facies progression from base to top is as follows: coarse-grained lag deposits of the thalweg; medium to fine-grained, festoon cross-stratified sandstone; fine-grained, planar-bedded sandstone; fine to very fine-grained, current ripple-laminated sandstone; and abandonment fill (sparsely burrowed siltstones with occasional climbing ripples). The thalweg deposits are well cemented resulting in non-reservoir rock. Early cementation occurred in the most permeable facies prior to oil migration. The lower energy facies, which were less permeable, were not affected by this phase of cementation.

Oil migration into the Spearfish is postulated to be secondary, from a breached Mississippian trap in the Starbuck Southwest field to the west, where the point bar sandstone is in contact with the underlying Charles Formation (Berentson Member) reservoir.

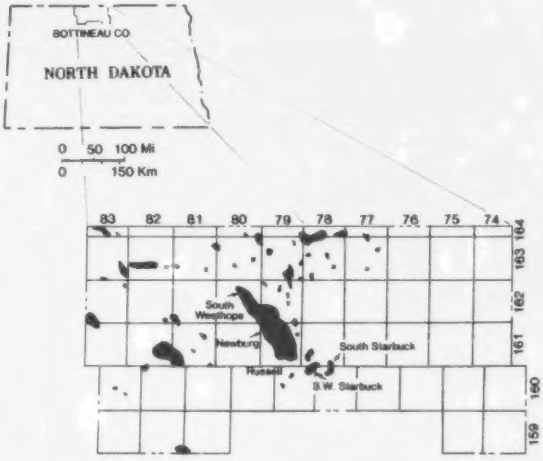
Lack of an active water drive has resulted in relatively low recoveries (40,000 bbl/well). Similar facies are productive in the Russell field (Township 160N, Range 79W) where recoveries are much higher (110,000 bbl/well), due to the presence of an excellent edge-water drive.

INTRODUCTION

The Triassic is petroleum productive on the eastern flank of the Williston Basin. Production is often commercial due to shallow drilling depths and relatively low completion costs. However, despite these incentives, exploration for Spearfish fields is rare. Spearfish "exploration" is generally relegated to serendipity while drilling Mississippian prospects. Perhaps a better understanding of the trapping and producing facies and of the petroleum migration patterns will aid in the formulation of Spearfish exploration strategies.

Bottineau County, North Dakota, has several fields (Fig. 1) which produce from the Triassic Spearfish Formation. Most notable are the Newburg (Twp. 161-162N, Rge. 79-80W) and South Westhope (Twp. 162N, Rge. 80W) fields which have yielded more than 38.8 Mbbls (through 1990), making them one of the largest Triassic stratigraphic reservoirs within the Williston Basin. Both fields have a natural waterdrive and wells that have produced in excess of 300,000 bbls from a depth of 3400 feet (LeFever and LeFever, 1991). The Newburg field has been unitized and is currently being waterflooded.

To the southeast and south are the smaller South Starbuck and Russell oilfields. Excellent core control in these two fields provides a basis for establishing the depositional framework of the reservoir facies.



Modified after Lefever and Lefever, 1991
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Figure 1: Location of oilfields on east flank of Williston Basin. South Westhope, Newburg, Russell and South Starbuck are productive from the Spearfish Formation (after LeFever and LeFever, 1991).

STRATIGRAPHIC SETTING

Zieglar (1955) subdivided the Spearfish Formation into the Belfield, Pine Salt and Saude Members (O-Y). Production within Bottineau County is from the Saude Member (Dow, 1959).

Dow (1964), in a regional study of the Spearfish, correlated the formation from the outcrops in South Dakota and Wyoming to the subsurface of the Williston Basin. He recognized the onlapping nature of the sequence which indirectly emphasized the significant basal lacuna. He also noted the role of the Spearfish on the east flank of the Williston Basin as a seal to the underlying Mission Canyon and Charles reservoirs. Dow also noted the presence of a basal unit which is "made up of sandstone, conglomerate, and locally contains fragments of chert and limestone derived from the pre-Spearfish surface". In describing this unit, the reservoir rocks at Newburg and South Westhope, Dow made no direct environmental interpretation but "considered them to be an onlapping transgressive unit".

Other depositional fabrics noted by earlier workers included recognition of a basal Watrous Formation (Canada) sandstone, consisting of an orange-red frosted grain unit of possible eolian origin (Milner, 1954). The presence of this frosted grain unit within the basal Saude, plus an abundance of frosted grains throughout the interbedded sandstone layers indicated an eolian depositional environment. A broad alluvial plain was inferred to exist across much of the basin during Saude time (Goldsmith, 1959).

DEPOSITIONAL INTERPRETATION

All cores available from wells in the South Starbuck and Russell fields were described to interpret the depositional facies and to gain a better understanding of the reservoir geometry. The reservoir rocks in both fields are interpreted to be point bar sandstones deposited by a meandering fluvial system.

The vertical sequence of bedforms varies depending upon location within the point bar. A nearly complete vertical sequence is present in the Turtle Mountain - 1-9 Brandt Goodman, se/nw, Sec. 9, 160N, 78W. Photographs of these facies depict from the base upwards coarse-grained to granule-sized channel lag deposits (Fig. 2), medium to fine-grained, festoon cross-bedded sandstone (Fig. 3), fine-grained, planar-bedded sandstone (Fig. 4), fine to very fine-grained current ripple-laminated sandstone (Fig. 5), and sparsely burrowed siltstones with occasional climbing ripples (Fig 6). This facies succession records waning depositional energy conditions. The base of the channel is in unconformity with either Mississippian carbonates (Fig. 7) or Spearfish burrowed mudstones similar to those overlying the fluvial deposits. Burrowing of the mudstones is extensive, although the reddish and greenish colour contrast allows the recognition of poorly preserved lamination (Fig. 8). These mudstones are interpreted to be marine in origin due to their highly burrowed character and regional consistency.

The depositional fabric of this fluvial sequence suggests that the most permeable facies was the high energy channel lag deposits, and that permeabilities decreased upward through the sequence. An early stage of cementation, prior to oil migration, thoroughly cemented the lag deposits. This high degree of cementation is a key to recognizing thalweg deposits on well logs, where the well-cemented rocks have very low porosities despite their clean gamma-ray character.

Figure 2: Coarse-grained to granule sized, channel lag deposits, Turtle Mtn.- 1-9 Brandt Goodman, ne/nw 9-160N-78W.

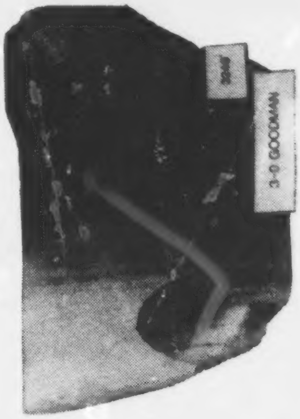

Figure 3: Medium to fine-grained, festoon cross-bedded sandstone, Turtle Mtn. - 3-9 Goodman, nnw nnw 9-160N-78W.

Figure 4: Fine-grained, horizontally laminated sandstone, Turtle Mtn. - 3-9 Goodman, nw/nw 9-160N-78W.

Figure 9 is a depositional dip cross-section across the channel at South Starbuck field. Three fining-upwards sequences can be recognized (denoted by time lines T_1, T_2, T_3). Within the time-bound units, facies changes occur both laterally and vertically. The southernmost well (nw/sw 10), K&E Petroleum - 1 Laudenschlager, which is located in the channel thalweg, contains only the coarse-grained, well-cemented thalweg deposits overlain by the abandonment fill facies. North of this well, grain size decreases and the Fundamental Oil-1 Solberg (sw/ne 9) well is productive from medium-grained festoon-cross bedded sandstones. The third well in the cross-section, Turtle Mountain - 1-9 Brandt Goodman, is located higher up on the point bar and displays both festoon cross-bedded sandstones and horizontal laminated facies (although the former is predominant). These beds are overlain by abandonment fill. The northernmost well, Turtle Mountain - 3-9 Goodman, represents the lowest depositional energy and consists of thin festoon cross-bedded, thick horizontal laminated and ripple-laminated sandstones and siltstones overlain by abandonment fill facies. The well developed festoon cross-bedded and horizontal facies are the hydrocarbon reservoirs in the field.

Figure 5: Fine to very fine-grained, current ripple-laminated sandstone, Turtle Mtn. - 3-9 Goodman, nw/nw 9-160N-78W.

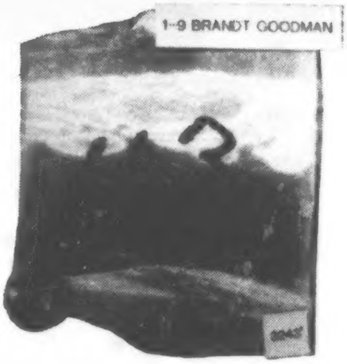

Figure 6: Sparsely burrowed siltstone with climbing ripples, Turtle Mtn. 1-9 Brandt Goodman, ne/nw 9-160N-78W.

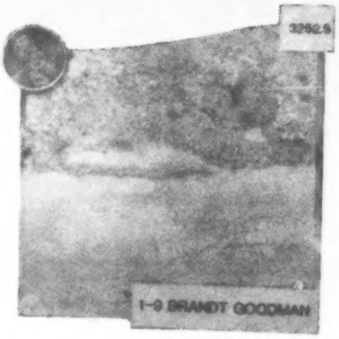

Figure 7: Contact between Spearfish (Saude Member) channel sandstone and the underlying Mississippian Charles Formation (note carbonate intraclasts in sandstone) Turtle Mtn. 1-9 Brandt Goodman ne/nw 9-160N-78W.

PETROLEUM GEOLOGY

The South Starbuck oil reservoir occurs on a homocline dipping to the southwest (Fig. 10). The known oil column is approximately 80 feet; however, the updip and downdip limits of the field are not defined. Trapping is confined to the area of the active channel fill. Figure 11 is an isopach map of the active channel fill within the point bar deposit. The crosshatched areas denote the location of the axis of the thalweg deposits. Recognition and mapping of this facies are important because they allow for the prediction of reservoir quality rock in the festoon and horizontal laminated sandstones. Note the sinuous trace of the thalweg deposits. This sinuosity suggests that the South Starbuck field consists of several coalesced point bar deposits.

Oil accumulations in the Spearfish are migrational from underlying Mississippian reservoirs (Folsom *et al.*, 1958). All four fields, Newburg, South Westhope, Russell and South Starbuck, overlie truncated Berentson or Midale Members of the Charles Formation (Anderson and Carlson, 1965). Where the Triassic rocks are mudstones, vertical migration is hindered and the possibility of trapping within the Charles is increased. This relationship has significant exploration significance. When drilling for Charles prospects it is important to have an impermeable top seal. Thus a good Charles prospect is a poor Spearfish prospect. Conversely the presence of an adjacent Spearfish sandstone is detrimental to a Charles prospect.

Recovery efficiencies within Spearfish reservoirs are often quite high for this shallow depth (Table 1). With a thick oil column and high oil saturations, the recovery efficiency seems to be primarily a function of maintaining high reservoir energy. Note the large discrepancy between the recoveries in the Russell and South Starbuck fields. These fields are both productive from point bar sandstones, exhibit similar oil saturation, permeability, and thickness, yet their production characteristics are very different. High recovery efficiency in the Russell field is associated with an active water drive and abundant water production. This contrasts with the lower volume, nearly water-free production from the South Starbuck field. Excellent recoveries from the Newburg and South Westhope fields are also aided by natural and secondary water drives. This discrepancy between water-drive and solution-gas drive reservoirs suggests that a waterflood at South Starbuck should be initiated.

Table 1: Cumulative Oil/Water Production (bbls) from Spearfish fields (1990).

FIELD	CUMULATIVE OIL PRODUCTION	CUMULATIVE WATER PRODUCTION
Newburg (Spearfish/Charles)	27,709,769	19,664,839
Russell	458,535	700,202
South Starbuck	118,657	302
South Westhope	11,188,354	6,985,495

Figure 8: Bioturbated marine mudstone Turtle Mtn. 1-9 Brandt Goodman, ne/nw 9-260N-78W

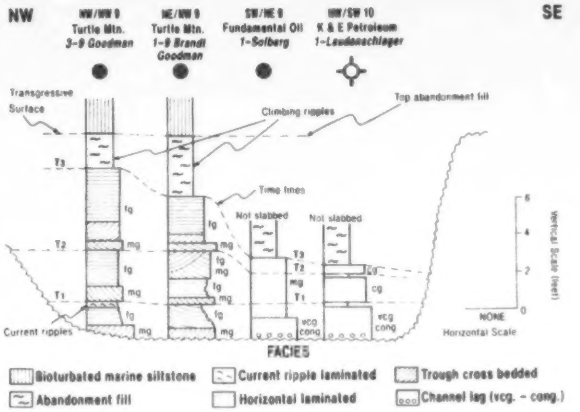

Figure 9: Dip cross-section of South Starbuck field showing three fining-upwards sequences. Within time-bound units, decreasing energy is demonstrated by both the vertical change in bedforms and also by laterally equivalent lower energy facies to the northwest.

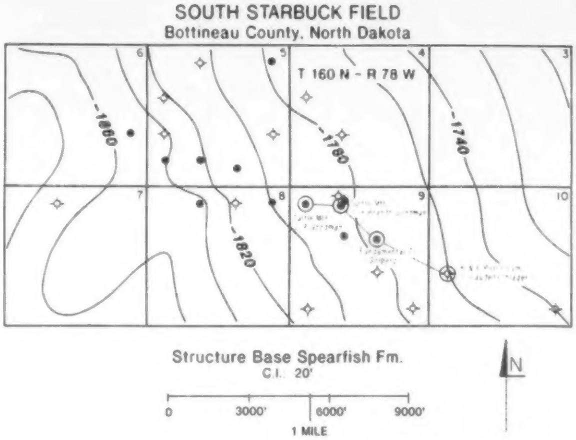

Figure 10: Structure contour map at base of the Spearfish Formation in the South Starbuck field.

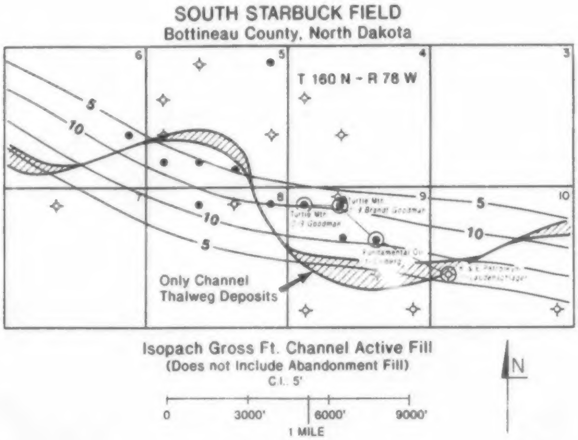
SOUTH STARBUCK FIELD
Bottineau County, North Dakota
T 160 N - R 78 W
6
5
4
3
7
8
9
10
5
10
Only Channel Thalweg Deposits
Isopach Gross Ft. Channel Active Fill
(Does not include Abandonment Fill)
C.I.: 5'
0
3000'
6000'
9000'
1 MILE
N

Figure 11: Isopach gross feet active channel fill in the South Starbuck field. Crosshatched area contains only thalweg deposits.

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DEPOSITIONAL HISTORY OF THE JURASSIC SYSTEM AND HYDROCARBON ACCUMULATIONS IN THE WAPELLA-MOOSOMIN AREA, SOUTHEASTERN SASKATCHEWAN

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ABSTRACT

The Triassic/Jurassic section of southeastern Saskatchewan is an unconformity bounded sequence between the Mississippian subcrop and the Cretaceous Mannville Formation. Four lithostratigraphic units are recognized in the Jurassic of southeastern Saskatchewan: the Watrous, Gravelbourg, Red Jacket and Rierdon formations. The Red Jacket is the facies equivalent to strata between the lower member of the Gravelbourg and the Rierdon Formation.

The Triassic/Jurassic(?) red beds of the lower member of the Watrous Formation represent the earliest post-Mississippian deposits in the study area. They appear to have been laid down in a restricted marine environment but with repeated exposure, oxidization, and reworking. This phase was followed in the Bajocian by less restricted marine deposition, represented by anhydrites, dolomites and dolomitic mudstones of the upper member of the Watrous Formation and dolomitic limestones (unit A) of the lower member of the Gravelbourg Formation. Interbedded clastics and carbonates (Unit B) of the lower Gravelbourg Formation were deposited in a near-normal marine environment. In contrast, clastics and subordinate carbonates of the succeeding Red Jacket Formation were deposited during the Bajocian and Bathonian in depositional environments ranging from freshwater to brackish-water to open marine. This variation was probably caused by minor yet frequent fluctuations in sea level, causing broad shifts in the depositional environments. The unstable period of Red Jacket deposition was followed by a major marine transgression and shales of the Rierdon Formation were laid down in the study area during the Bathonian to Callovian.

Most Jurassic hydrocarbon production in southeast Saskatchewan is from the Rocanville and Moosomin members of the Red Jacket Formation in the Wapella-Moosomin area. Oil shows are also in the lower member of the Gravelbourg Formation; however, only one well is presently producing from this unit.

The dissolution of salt beds from the Middle Devonian Prairie Formation appears to have played a significant role in the depositional and hydrocarbon accumulation history of the Wapella-Moosomin area.

INTRODUCTION

The Jurassic sediments of the Wapella-Moosomin area southeastern Saskatchewan (Fig. 1) were studied in detail by Kreis (1991). The highly intriguing succession of sandstones, shales, carbonates and evaporites reviewed in this paper comprises the Watrous Formation, lower member of the Gravelbourg Formation, Red Jacket Formation and the Rierdon Formation (Fig. 2). The focus, however, is on the hydrocarbon-bearing sediments of the Red Jacket Formation.

Medium gravity oil is produced from discontinuous, unconsolidated to variably cemented sands of the Middle Jurassic Moosomin and Rocanville members of the Red Jacket Formation in the Wapella, Coothill and Red Jacket pools. In addition to these defined pools, there is significant Middle Jurassic production from the North Wapella and Moosomin localities. Hydrocarbon reserves are also present in a few wells proximal to the aforementioned locations. Within the Wapella field there is also comingled production of medium gravity oil from a basal sandstone unit (*i.e.* Wapella Sandstone) of the Cretaceous Mannville Formation. Oil is also produced in the Rocanville pool from the Mississippian Bakken Formation.

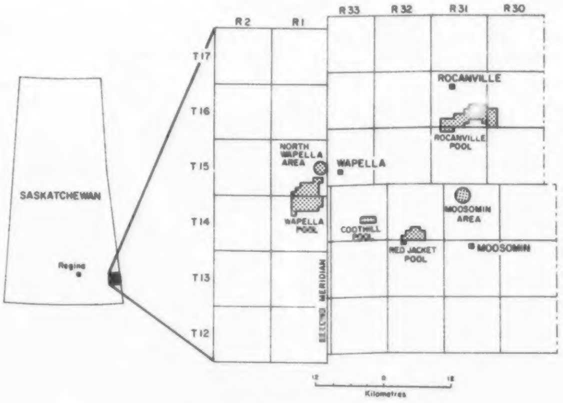


Figure 1: Location map of Wapella-Moosomin area.

GEOLOGICAL SETTING

The intracratonic Williston Basin was an actively subsiding structure during much of the Phanerozoic, with the greatest paleobathymetric control on sedimentation occurring during Silurian, Mississippian and Jurassic time (Kent, 1987). Jurassic strata in the Wapella-Moosomin area were deposited on the eastern flank of this intracratonic depression (Fig. 3) and form a broad sheet concordant to slightly disconcordant with respect to the underlying erosional surface on the Mississippian (Kreis, 1991). The present-day structural configuration of the Williston Basin is believed to be largely due to Laramide tectonics (Christopher *et al.*, 1971).

STRATIGRAPHY

The Triassic/Jurassic section of southeastern Saskatchewan is an unconformity bounded sequence between the Mississippian

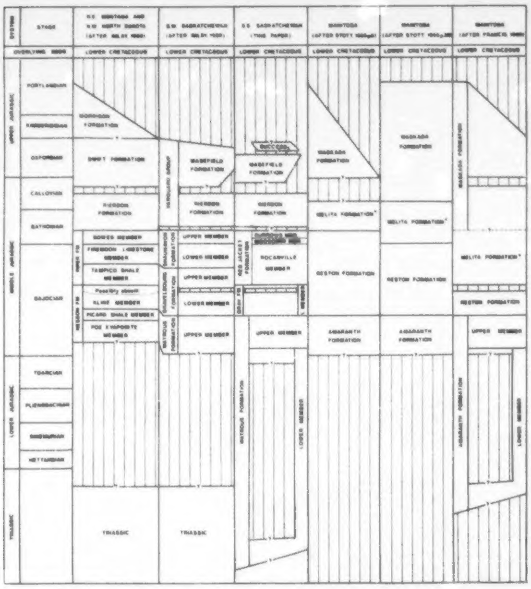


Figure 2: Stratigraphic correlation chart of Jurassic units in southeastern Saskatchewan and adjacent areas.

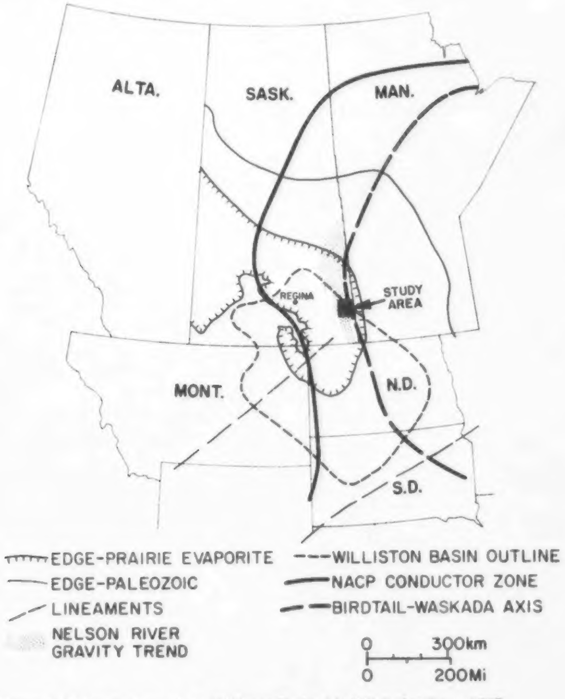


Figure 3: Major features of the Williston Basin. Modified after Kent (1987).

subcrop and the Cretaceous Mannville Formation. Five major mappable Jurassic units are recognized in the subsurface in the Wapella-Moosomin area. They are, in ascending order, the Triassic/Jurassic(?) red beds of the lower member of the Watrous Formation; the Middle Jurassic upper member of the Watrous Formation, the lower member of the Gravelbourg Formation, and the Red Jacket Formation; and the Upper Jurassic Rierdon Formation (Fig. 2).

Much of the Middle Jurassic section in southeastern Saskatchewan defies detailed correlation with the type area in southwestern Saskatchewan because of facies changes, numerous discontinuities, and a general lack of regionally persistent stratigraphic markers. The name Red Jacket Formation was proposed by Kreis (1991) for strata in southeastern Saskatchewan equivalent to the Shaunavon Formation and the upper member of the Gravelbourg Formation of the type area in southwestern Saskatchewan (Fig. 2). An arbitrary lithofacies cut-off in southeastern Saskatchewan was chosen to separate the Red Jacket Formation from those units (Figs. 4 and 5).

The type section for the Red Jacket Formation is found fully cored in the Tide Water North Wapella Crown 14-22-15-1 well (Lsd. 14-22-15-1W2) between 665.4 to 734.6 m depth (2183-2410 ft) (Fig. 6). The Red Jacket Formation is formally divided into three members. They are from oldest to youngest the Rocanville, Moosomin and Burrows. The divisions are dependent on recognition of a number of lithological changes and discontinuities chosen as unit boundaries. Each member is further divided into the following units: Rocanville A, B, and C; Moosomin A and B; and Burrows A and B. In addition, Moosomin A is subdivided in places into A1 and A2. Rocanville A and Rocanville C are not present in the type section of the Red Jacket Formation but are found in the reference Tide Water Wapella 6-33-14-1 well (Lsd. 6-33-14-1W2) (Fig. 7).

Lithological, palynological and stratigraphic evidence suggest that Rocanville A is equivalent to the upper member of the Gravelbourg of southwestern Saskatchewan and is Bajocian. Similar lines of evidence suggests Rocanville B and Rocanville C are equivalent to the Bajocian Lower Shaunavon of southwestern Saskatchewan and the Moosomin and Burrows members to the Bathonian Upper Shaunavon (Fig. 2).

DEPOSITIONAL HISTORY

The relative paucity and highly uneven distribution of boreholes and cores in the Wapella-Moosomin area preclude the

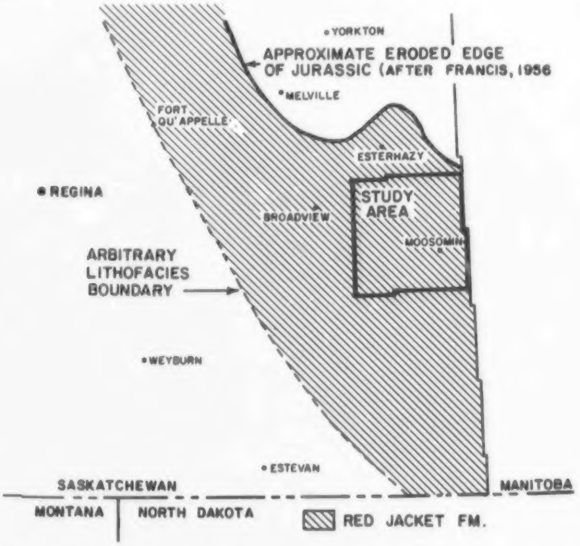


Figure 4: Location map showing distribution of Red Jacket Formation in southeastern Saskatchewan.

construction of detailed sedimentological models for much of the strata investigated. Nevertheless, sufficient information is available for a broad reconstruction of the paleo-environments and depositional settings.

WATROUS FORMATION

Lower Watrous

A lack of fossils in the lower Watrous red beds prohibits the assignment of an unequivocal age for this unit; however, stratigraphic evidence suggests that it is largely Early Triassic with an undefined portion of the uppermost part possibly being Early to Middle Jurassic.

The lower Watrous is characterized by moderately reddish-brown to dark reddish-brown, dolomitic to anhydritic siltstone and mudstone, with rare lenses of poorly sorted fine- to medium-grained quartzose sandstone near the base. Quartz grains are commonly pitted and frosted, and are often found floating in a siltstone or mudstone matrix.

The regional continuity of the lithological and geophysical log characteristics of this unit suggests that it was, at least in part, deposited in a marine environment (Barchyn, 1982; Christopher, 1984). Palynological work by Pocock (1972) supports this, but also indicates, that these sediments were likely proximal to a coastal desert.

The anhydritic nature of this unit attests to an evaporitic depositional environment, and the ubiquitous reddish-brown coloration strongly suggests periodic subaerial exposure and extensive oxidization. The common occurrence of pitted, frosted and floating grains indicates wind transport from nearby aeolian deposits.

Upper Watrous

The upper Watrous oversteps the lower Watrous towards the northeast in the Wapella-Moosomin area (Fig. 8), suggesting an outward expansion of the Williston Basin during the Bajocian. The presence of interbedded anhydrite, dolomite and dolomitic mudstone clearly indicates a restricted marine setting. Palynological evidence reported by Pocock (1972) supports this interpretation.

The presence of thin beds of fine-grained quartzose sandstone and scattered grains of clear quartz near the top of the upper Watrous in some wells reflects proximity to one or more terrestrial sources in southeastern Saskatchewan.

GRAVELBOURG FORMATION

Lower Gravelbourg

The interbedded anhydrite, dolomite and dolomitic mudstones of the upper Watrous grade up into dolomites and argillaceous dolomitic limestones of unit A of the lower Gravelbourg. The evaporitic upper Watrous increases in thickness at the expense of unit A of the lower Gravelbourg in southeastern Saskatchewan. This implies that evaporitic conditions persisted in southeastern Saskatchewan while coeval dolomitic limestones of unit A were being deposited farther to the west. Locally, unit A is an interbedded oolite and algal-laminated dolomitic limestone, suggestive of deposition in a shallow-water but less restricted marine setting than that of the upper Watrous.

Unit A is terminated at a disconformity in both southwestern and southeastern Saskatchewan. This disconformity is characterized by a weathered zone of dolomite with abundant chalcedonic chert, and rare euhedral bipyramidal crystals of secondary quartz. It is likely that unit A overstepped upper Watrous sediments northeasterly across the study area, but was eroded back to its present limit (Fig. 8).

In the Wapella-Moosomin area, interbedded calcareous mudstone and argillaceous limestone of unit B of the lower Gravelbourg Formation overstep unit A in a northeasterly direction (Fig. 8). The oolitic, peloidal, bioclastic and glauconitic nature of some of the limestones in this unit strongly suggests that these rocks were originally deposited in a marine environment. Some bioclastic limestone interbeds develop into anomalously thick (15 m) discontinuous bodies showing graded bedding, low- to high-angle cross-stratification and numerous truncation surfaces (Plate 1A). These bodies are interpreted to be near-shore bars and shoals. A dominantly marine setting is supported by an abundance of microforaminifera in this unit (Kybett and Nambudiri, 1989).

The presence of thin-shelled pelecypods (*Modiolus* ?), and a few specimens of typical freshwater species of ostracodes like *Limnocythere* and *Darwinula* and the brackish-water *Norcanollela* in some wells suggests local influence by waters of lesser salinities. Perhaps dilution of these waters by rivers with a high rate of discharge, or deposition in embayments with poor communication to normal marine waters, or both, may explain these occurrences. It is also possible that these ostracodes were transported into the marine environment by rivers and redistributed by currents.

RED JACKET FORMATION

Rocanville Member

Lithological and palynological evidence from Rocanville A indicates a rapid change-over from the marine environment of unit B of the lower Gravelbourg to a marginal-marine to continental depositional environment. The contact between these units is sharply delineated in core but is less obvious on geophysical logs in some locations owing to their similarity in log responses. Rocanville A is composed of interbedded and interlaminated very fine- to fine-grained sandstone and sandy, noncalcareous, micaceous and carbonaceous siltstone and minor mudstone. Palynological evidence indicates an inland coastal plain bordered by extensive swamps (Pocock, 1972).

A disconformity separates Rocanville A and Rocanville B in the Wapella-Moosomin area. Rocanville A pinches out toward the north and east such that within the Wapella-Moosomin area it is recognized in only a few wells (Fig. 5) and is less than 1.5 m thick. It is likely that widespread pre-Rocanville B erosion also contributed significantly to the removal of Rocanville A in this area.

Generally, stacked sheet-like bodies of sandstone characterize Rocanville B. These sandstones are very fine- to fine-grained, moderately sorted to well sorted, subangular to subrounded, clear quartz. They blanket most of the study area and become thicker toward the north and east (Figs. 9 and 10), possibly indicating proximity to and direction a sand source.

Glauconitized mica flakes in some of these sheet-like sandstone bodies indicate a marine origin. This deduction is supported by the presence of marine palynomorphs, carbonate interbeds, fish remains, pelecypods and rare gastropods. The relative abundance of terrestrial palynomorphs compared to marine palynomorphs in Rocanville B, however, suggests a near-shore marine setting following Traverse's (1988) suggestion that spores and pollens are present in virtually all depositional environments but that their relative abundance decreases towards

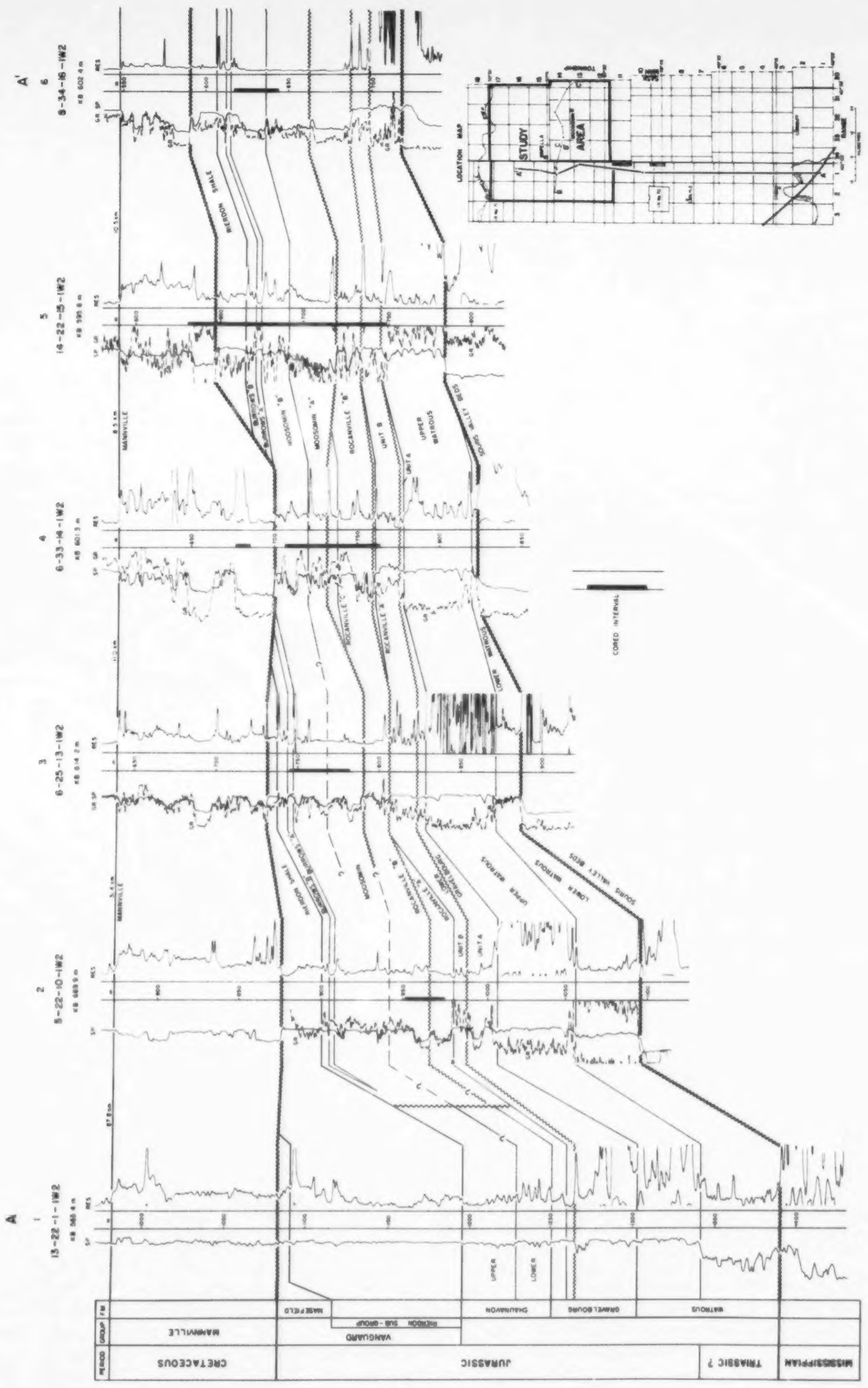


Figure 5: North-south stratigraphic cross-section A'-A of Jurassic strata in southeastern Saskatchewan.

the open ocean.

The low-angle planar cross-stratification, small-scale current ripples and massive bedforms, together with the widespread distribution of the basal sandstones in Rocanville B suggest that they may be stacked near-shore marine shelf sands and possibly relict beach deposits. The sharp nature of the contacts between interbeds in these sandstones indicates rapid shifts in energy conditions - conditions one would expect in such an environment.

The presence of channel sandstone (Plate 1B) at the base of Rocanville B in the Pondex Pat Parkman 5-22-10-1W2 well immediately south of the Wapella-Moosomin area (Fig. 5), suggests that channel sandstones may have locally infilled the incised surface below the base of Rocanville B. The channel facies in this well grades up into sandstones identical to the sheet-like sandstone bodies at the base of Rocanville B in the study area.

The upper half of Rocanville B is characterized by variegated dolomitic mudstones, very fine- to fine-grained argillaceous and glauconitic/chamositic(?) sandstones, siltstones, and minor interbeds of argillaceous dolostone. The variegation of the rocks (Plate 1C) ranges from the dark reddish-brown through greyish-orange indicative of oxidizing conditions (Berner, 1971). This, dolomitization, and the presence of desiccation cracks (Plate 1C) indicate subaerial exposure at times and perhaps deposition in a supratidal hypersaline environment.

Freshwater ostracodes intermixed with marine palynomorphs and glauconite in this unit may be explained by periodic oscillation of the strandline on a shelf of very low gradient. Even minor fluctuations in sea level would cause widespread subaerial exposure or flooding and result in considerable reworking. Periodically exposed restricted marine to supratidal depositional environments, on a shelf of low gradient, are therefore envisaged for the upper half of Rocanville B.

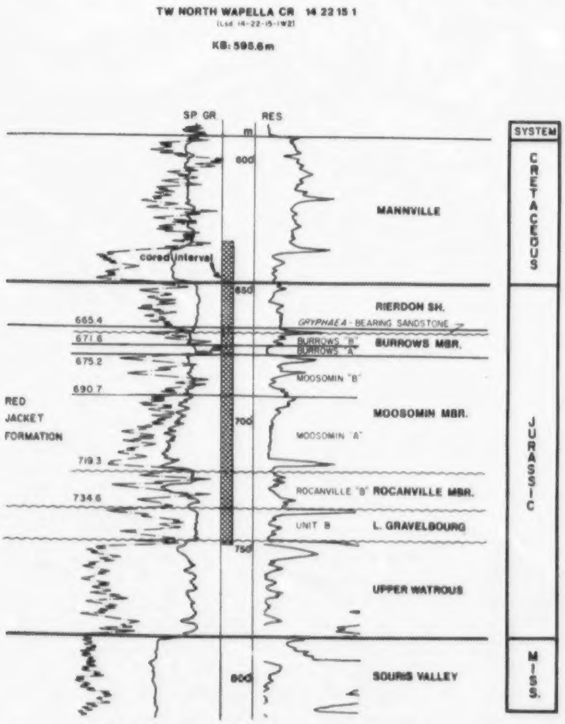


Figure 6: TW North Wapella Cr 14-22-15-1W2 well showing cored interval through type section of the Red Jacket Formation and its members. Note: Units Rocanville A and Rocanville C of the Rocanville Member are not present.

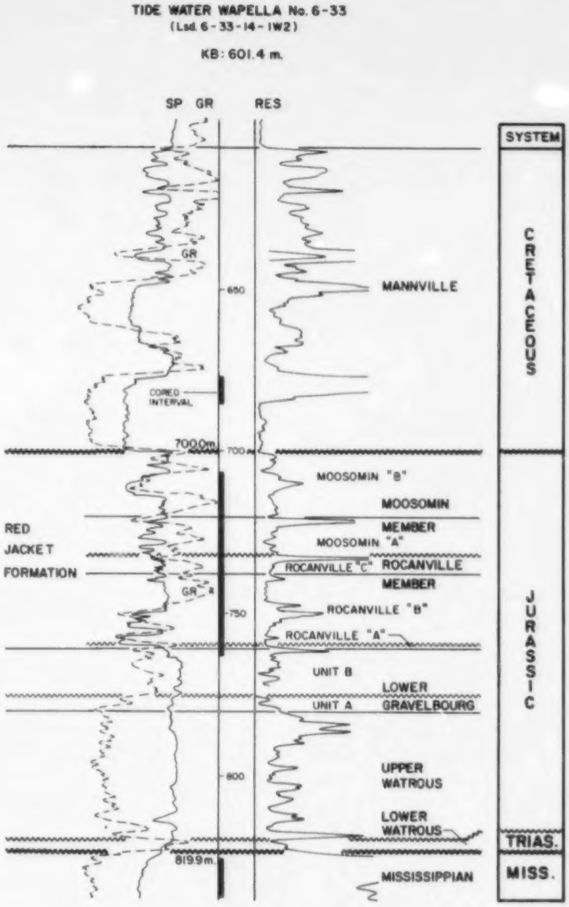


Figure 7: Tide Water Wapella No 6-33 well showing cored interval through reference section for the Rocanville Member.

The contact between Rocanville B and Rocanville C is sharp and is considered to represent a diastem, an interpretation supported by an apparent caliche with calcite-infilled rootlet molds(?) at the top of Rocanville B in the CNW Wapella 14-2-15-1W2 well. Rocanville C is made up of interbedded sandstone, siltstone, mudstone, and oolitic bioclastic limestone, with interlaminated sandstone and siltstone being dominant. Glauconitic pellet-like masses, dinoflagellates, marine ostracodes, pelecypods and, above all crinoids and corals demonstrate that Rocanville C was deposited in a dominantly open marine environment. Thus Rocanville B deposition was followed immediately by a marine transgression.

Moosomin Member

An erosional disconformity separates the Moosomin and Rocanville members. The Rocanville Member is eroded to a stratigraphic depth of 30 m in some areas. The subunit Moosomin A1 infills paleo-valley and lowland sites on the Rocanville paleotopographic surface (see Figs. 9, 10 and 11). Moosomin A1 is an interbedded and interlaminated sequence of carbonaceous and pyritous sandstone, siltstone and mudstone. The olive-grey to medium grey colour, presence of disseminated pyrite and nodular pyrite and abundance of well-preserved carbonaceous detritus in this subunit suggests a reducing depositional environment (Christopher, 1964).

Several lines of evidence also point towards a brackish-water or estuarine depositional setting for Moosomin A1. These include: a low diversity but locally rich ichnofossil suite composed solely of *Planolites* and *Chondrites* (Ekdale *et al.*, 1984), the coexistence of pyrite and siderite (Wightman *et al.*, 1987), and abundant terrestrial palynomorphs. The presence of sandstones showing bi-directional cross-bedding suggests opposing flow directions (Plate 1D) by tidal processes as may be found in an estuary (Clifton, 1982).

On the basis of paleogeographical and sedimentological evidence, and the brackish water character of Moosomin A1, the Wapella-North Wapella and Red Jacket-Rocanville paleo-valleys are interpreted to be estuaries, and the lowland areas are interpreted to be coastal embayments (Fig. 11). This follows Pritchard's (1967) definition that an estuary is a semi-enclosed coastal body of water which has free connection with the open sea but in which freshwater from the land measurably dilutes sea water.

The rare occurrence of marine palynomorphs like *Leiosphaeridia* and *Comasphaeridium*(?) sp. in this unit suggests brief incursions of salt water. Variations between marine and fluvial influences in these deposits resulted from minor fluctuations in sea level which cause an increase or decrease in the size of the tidal prism (Clifton, 1982). It is likely that the sediments in these paleo-valleys are dominantly estuarine.

The subunit Moosomin A2 conformably overlies Moosomin A1 in the paleo-valleys and lowland sites, but oversteps the paleohighlands on the eroded Rocanville Member (see Figs. 9, 10, and 11). This suggests that Moosomin A1 deposition was followed by a relative rise in sea level and deposition of Moosomin A2 proceeded in a more marine environment. Lithological and paleontological evidence support this interpretation. Although Moosomin A2 is lithologically similar to, and in gradational contact with Moosomin A1, it differs from Moosomin A1 by being less carbonaceous, by minor interbeds of bioclastic limestone near its top, and by its glauconitic sandstones. Its few species of dinoflagellates, acritarchs and foraminifera also support a marine interpretation. More specifically, the abundance of *Ammodiscus* suggests a marine environment of lowered salinities, as may be found in a lagoonal setting sheltered behind a barrier-island complex and diluted by freshwater runoff (Braun, 1989, pers. comm.). The presence of the miliolid *Opthalmidium* also indicates a low-energy near-shore depositional environment (Brooke and Braun, 1972).

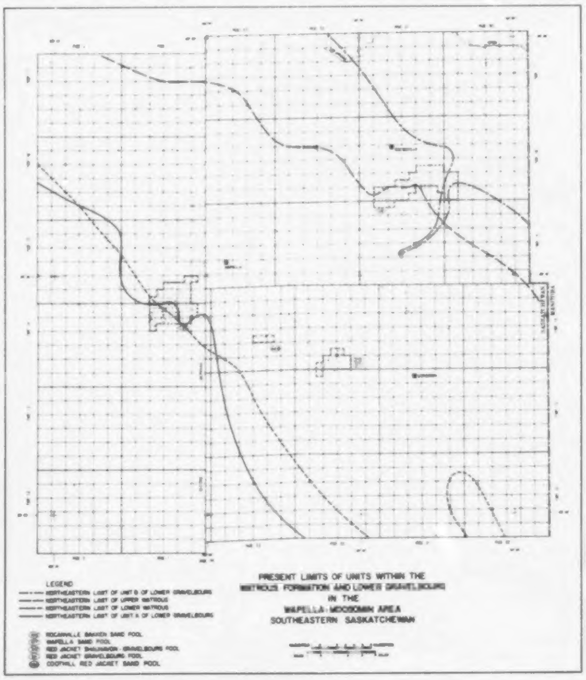

LEGEND
NORTHEASTERN LIMIT OF UNIT B OF LOWER GRAVELBOURG
NORTHEASTERN LIMIT OF UPPER WATROUS
NORTHEASTERN LIMIT OF LOWER WATROUS
NORTHEASTERN LIMIT OF UNIT A OF LOWER GRAVELBOURG
ROCANVILLE BAKKEN SAND POOL
WAPELLA SAND POOL
RED JACKET SHAUNAVON - GRAVELBOURG POOL
RED JACKET GRAVELBOURG POOL
COOTHILL RED JACKET SAND POOL
PRESENT LIMITS OF UNITS WITHIN THE
WATROUS FORMATION AND LOWER GRAVELBOURG
IN THE
WAPELLA-MOOSOMIN AREA
SOUTHEASTERN SASKATCHEWAN

Figure 8: Present limits of units within the Watrous Formation and lower Gravelbourg in the Wapella-Moosomin area in southeastern Saskatchewan.

Sandstones in Moosomin A2 commonly show flaser and wavy bedding, typical of tidal (Reineck and Wunderlich, 1968) and subtidal facies (Weimer *et al.*, 1982). Many of these sandstones are well sorted and may reflect a barrier-beach or tidal delta setting.

The contact between Moosomin A2 and Moosomin B is sharp in cores from the Wapella pool but shows little evidence of erosion. At the top of Moosomin A is commonly a dolomite- or calcite-cemented sandstone which is generally overlain by a thin bed of variegated (*i.e.* oxidized) mudstone. These characteristics indicate brief exposure and suggest a diastemic break between Moosomin A and B. The contact is rarely cored outside the Wapella pool area, and is not readily recognized on geophysical well logs, which make correlation highly tenuous beyond this area (Figs. 9 and 10).

Moosomin B comprises interbedded and interlaminated sandstone and greenish-grey mudstone with rare argillaceous limestone in the lower half. The dominantly greenish-grey colour and lack of carbonaceous detritus in the mudstones and siltstones of Moosomin B contrast with the light olive-grey to olive-grey mudstones and siltstones of Moosomin A2 and are interpreted to indicate deposition in a more oxygenated depositional environment (Potter *et al.*, 1980) than that envisaged for Moosomin A2. The greater abundance of marine palynomorphs, pelecypods, fish remains and ichnofossils in Moosomin B supports this hypothesis.

Flaser, wavy and lenticular bedding, particularly in the upper 10 m of Moosomin B, indicate repeated alternation between current or wave action and slack water (Reineck and Wunderlich, 1968). These bedding types are commonly associated with tidal rhythms and are considered important criteria in the recognition of a tidally influenced deposit (Weimer *et al.*, 1982; Reading, 1986).

Bioturbation is also common in Moosomin B, particularly in the upper 10 m of the unit where it commonly imparts a burrow-mottled texture to the sands and muds (Plate 1E). The low diversity but abundant trace fossils belonging to *Glossifungites* (firm substrate) and *Skolithos* (shifting substrate) ichnofacies (Plates 1F and 1G) indicates deposition in a biologically rigorous environment (Ekdale *et al.*, 1984). A freshwater to brackish-water ostracode assemblage and charophyte oogonia are evidence of a freshwater to brackish-water setting. Numerous truncation surfaces and escape burrows (Plate 1G) indicate extremes in current velocities resulting in erosion and rapid deposition. Syneresis cracks (Plate 1H) also suggest fluctuations in salinities, and mud cracks at the top of the unit indicate subaerial exposure (Plate 1I).

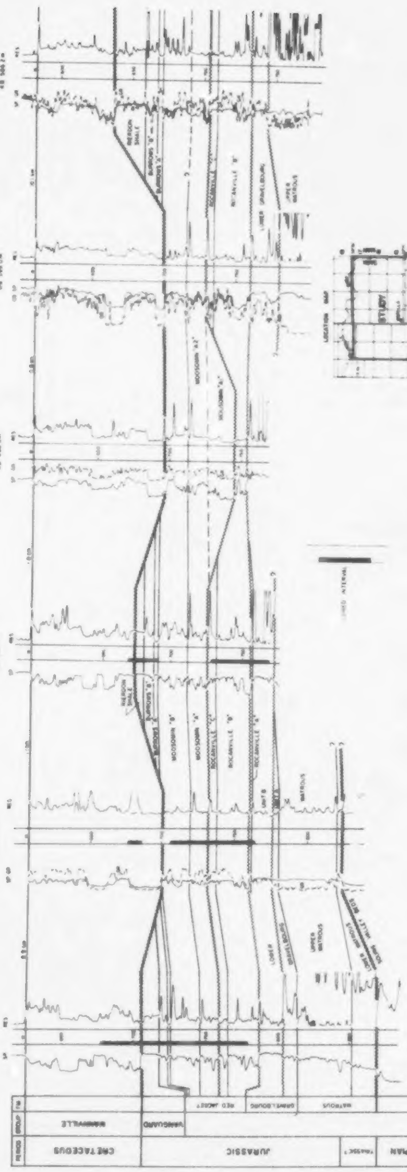


Figure 9: West-east stratigraphic cross-section B-B' of Jurassic strata in Wapella-Moosomin area, southeastern Saskatchewan.

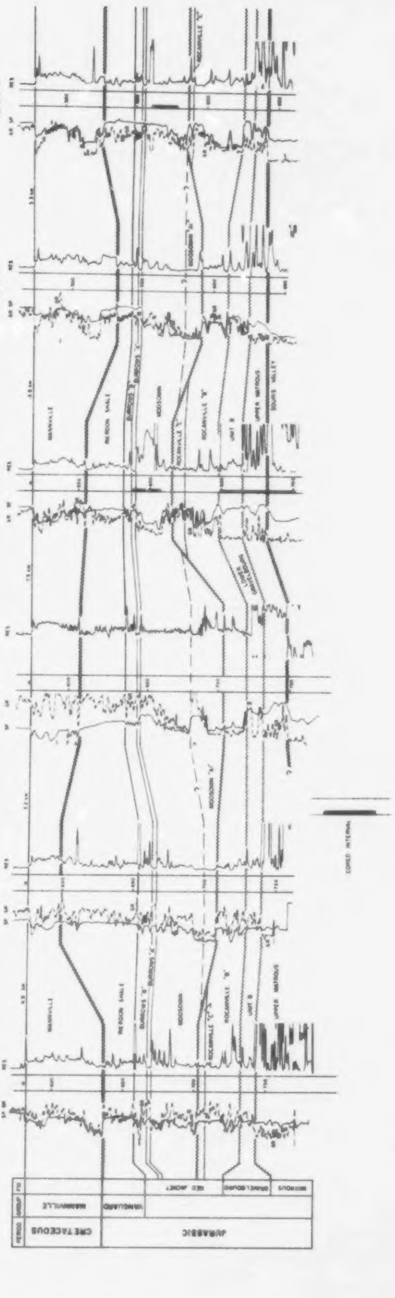


Figure 10: West-east stratigraphic cross-section C-C' of Jurassic strata in Wapella-Moosomin area, southeastern Saskatchewan.

Below this upper zone, Moosomin B is characterized by bioturbated interbedded sandstone and mudstone, and minor marlstone. These beds feature cut-and-fill very fine- to fine-grained, well-sorted fining upward sandstones up to 15 m thick. These sandstones are characterized by low- to high-angle planar (up to 30°) cross-stratification, and flaser and small-scale current bedding. Mudstone drapes are common (Plate 1J), as are mudstone pebbles and clasts at the bottom of these sandstones. Organic detritus indicating periods of slack water deposition commonly defines laminae in these sandstones and bioturbation is rare.

In view of the above evidence, a tidal flat depositional

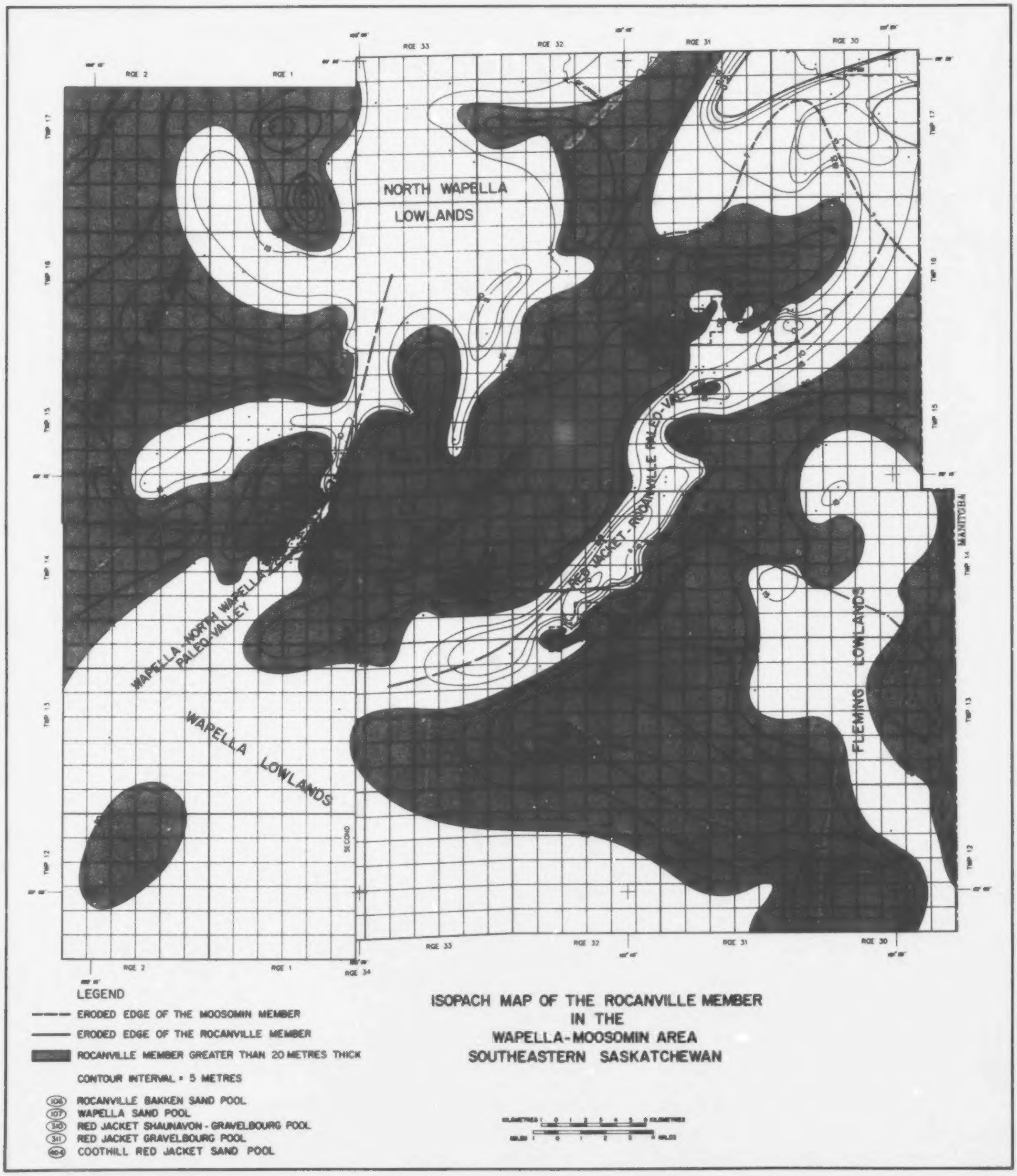

RGE 33
RGE 32
RGE 31
RGE 30
RGE 2
RGE 1
RGE 34
TWP 17
TWP 16
TWP 15
TWP 14
TWP 13
TWP 12
NORTH WAPELLA LOWLANDS
WAPELLA-NORTH WAPELLA PALEO-VALLEY
WAPELLA LOWLANDS
RED JACKET - ROCANVILLE PALEO-VALLEY
FLEMING LOWLANDS
MANITOBA
SECOND
LEGEND
ERODED EDGE OF THE MOOSOMIN MEMBER
ERODED EDGE OF THE ROCANVILLE MEMBER
ROCANVILLE MEMBER GREATER THAN 20 METRES THICK
CONTOUR INTERVAL = 5 METRES
(106) ROCANVILLE BAKKEN SAND POOL
(107) WAPELLA SAND POOL
(310) RED JACKET SHAUNAVON - GRAVELBOURG POOL
(311) RED JACKET GRAVELBOURG POOL
(404) COOTHILL RED JACKET SAND POOL
ISOPACH MAP OF THE ROCANVILLE MEMBER
IN THE
WAPELLA-MOOSOMIN AREA
SOUTHEASTERN SASKATCHEWAN
KILOMETRES
MILES

Figure 11: Isopach map of Rocanville Member showing paleogeographical features in the Wapella-Moosomin area, southeastern Saskatchewan.

environment is envisaged for Moosomin B. Tidal flats are biologically stressful, subject to extremes in currents, water depth, salinity, temperature, dessication, erosion and rapid deposition (Weimer *et al*., 1982). They are commonly divided into intertidal and subtidal environments. The sediments in the upper 10 m of Moosomin B are interpreted to be mostly intertidal and the sandstones to be infills of tidal channels in the subtidal portion of the tidal flat.

The overall fining upwards of bioturbated sands and muds, ending in the dessicated top of Moosomin B, suggests progradation of intertidal flat deposits over subtidal channel fill and tidal point bar deposits (Reineck, 1967; Klein, 1971; Evans, 1975).

BURROWS MEMBER

Burrows A immediately overlies Moosomin B and infills mud cracks in its dessicated surface (Plate 11). This unit is a medium grey to dark grey, weakly calcareous claystone and mudstone with minor interlaminations of siltstone. Minor layers of thin (2-3 mm) laminae containing abundant, small (less than 1 cm thick) molds of what are tentatively identified as the pelecypods *Pleuromya simplex* (?) and *P. oblongata* (?) also characterize this unit.

The presence of foraminifera like *Ammodiscus* sp. and *Opthalmidium* sp. in abundance suggests deposition in a relatively low energy, brackish-water to marine environment (Brooke and Braun, 1972). Gonyaulacacean dinoflagellates are also scattered throughout indicating deposition in an open marine depositional environment (Williams, 1978) and implying that a minor marine transgression occurred during Burrows A time. The uniform thickness of this unit and its widespread distribution in southeastern Saskatchewan supports this interpretation.

Burrows A grades rapidly up into lenticularly bedded and bioturbated sandstones and greenish-grey mudstones and siltstones of the overlying Burrows B which exhibits all of the sedimentological, paleontological and ichnological features of the upper 10 m of Moosomin B. This implies that the minor marine transgression of Burrows A was followed by a minor regression and a return to deposition on a similar tidal flat.

RIERDON FORMATION

The Rierdon is the youngest Jurassic formation in the Wapella-Moosomin area. It is unconformably overlain by the Cretaceous Mannville Formation throughout this area (Figs. 5, 9, and 10).

Lithological, paleontological and palynological evidence clearly demonstrate a marine depositional environment for the Rierdon Formation. In the Wapella-Moosomin area, a thin (less than 1.5 m) *Gryphaea*-bearing sandstone subunit commonly occurs at the top of Burrows B. In core, this subunit is found to be in sharp and erosive contact. It is interpreted to be a lag deposit marking the beginning of the major Rierdon marine transgression and deposition of calcareous open marine mudstones. It is difficult to distinguish between this subunit and sandstones at the top of Burrows B on the basis of geophysical logs. This subunit is therefore included in Burrows B for practical reasons (Figs. 6 and 7).

Figure 12 summarizes Jurassic transgressive/regressive events in the Wapella-Moosomin area.

PALEOTECTONIC CONTROLS ON SEDIMENTATION

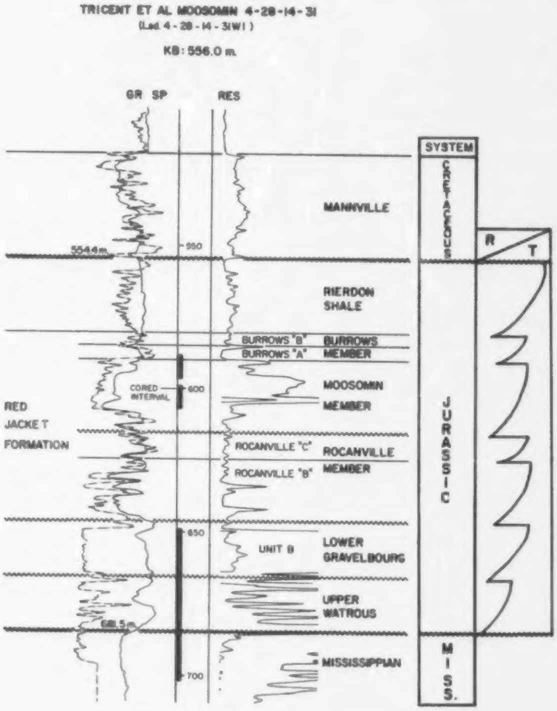


Figure 12: Tricent *et al*. Moosomin 4-28-14-31W1 well showing the occurrence of recognized transgressions (T) and regressions (R) in the Jurassic stratigraphic succession. Note that Rocanville A of the Rocanville Member is not present.

Kent (1987) described a number of structural elements which appear to indicate long-term weaknesses that may have had an effect on subsidence in the Williston Basin. The western portion of the basin is underlain by Archean basement of the northwestern Churchill and Wyoming structural provinces which, in turn, are interpreted to be components of an early Proterozoic paleocontinent. The eastern margin of this paleocontinent appears to be approximately coincident with the North American central plains (NACP) electrical conductivity zone (Fig. 3). The Birdtail-Waskada Axis in Manitoba is a north-south-trending geophysical feature consisting of numerous structural and isopach anomalies in Devonian and younger strata, and is considered to represent the approximate western margin of the Superior structural province (McCabe, 1967 and 1971). Lewry (1984) postulated the existence of a suspect terrane collage of early Proterozoic magmatic arcs, volcanogenic metasedimentary rocks, ultramafic rocks, and possible microcontinental remnants between the Churchill and Wyoming paleocontinents and the Superior craton.

The Birdtail-Waskada Axis is important to the present study as it passes directly through the Wapella-Moosomin area. The Nelson River gravity trend, which also passes through the area, is located immediately west of, and parallel to, the Birdtail-Waskada Axis (Fig. 3). This gravity trend may mark the boundary between the Churchill and Superior provinces (Macdonald and Broughton, 1980).

Christopher (1984) described a south-trending ridge of subcropping Mississippian strata along Longitude 102°W which separates the Saskatchewan portion of the Williston Basin (lower Watrous sub-basin) from the lower Amaranth sub-basin in Manitoba. He concluded that uplift of the intervening area took place sometime during post-Mississippian to pre-Triassic time. The uplifted area appears to be coincident with the Birdtail-Waskada Axis and the Nelson River gravity trend, and includes the Wapella-Moosomin area (Fig. 3). Kreis (1991) suggested that this uplifted area is at least in part related to reactivation and movement along an ancient suture zone separating the Churchill and Superior Provinces and be named the Birdtail-Waskada Arch.

Within the Wapella-Moosomin area the near coincidence of zero limits of the lower and upper members of the Watrous Formation, and units A and B of the lower member of the Gravelbourg suggests that there must have been some paleotectonic control on their present distribution. These units show a progressive overstepping of younger over older units in a northeasterly direction as they onlap a south-trending ridge of the Mississippian subcrop. The notable exception to this progression is unit A of the lower Gravelbourg, which can be explained by erosion of the top of the unit.

It is possible that truncation of some Middle Jurassic units in the study area may have been caused by uplift, as well as by contemporaneous regression (for example, truncation of unit A of the lower Gravelbourg and the Rocanville A unit (Fig. 5)). Similarly, the incision of up to 30 m below the unconformity on the Rocanville Member in the Wapella-Moosomin region (Figs. 9 and 10) might reflect: a) uplift of the study area or of the hinterland to the northeast, b) downcutting by rivers in response to lowered base levels, coincident with a relative drop in sea level (*i.e.* regressions), or c) a combination of regression and uplift.

HYDROCARBON ACCUMULATIONS

The Wapella-Moosomin area includes all known economic occurrences of hydrocarbons in Middle Jurassic reservoirs in southeastern Saskatchewan. Most hydrocarbon production in this area comes from the Wapella, Coothill and Red Jacket pools and a number of scattered wells nearby. Other production is taken from the North Wapella and Moosomin localities (Fig. 1). The oil produced is medium gravity, ranging from 906-960 kg/m^3 to (16°-25° API), from reservoirs at depths between 650-760 m. The shallow depth of burial has resulted in minimal compaction of reservoir sandstones such that porosity is generally greater than 25 percent.

The Wapella pool has been in production since 1952 from a basal Mannville (Lower Cretaceous) Wapella sandstone (Dina equivalent?) and from a number of discontinuous sandstone bodies in the Moosomin and Rocanville members. Hydrocarbon production from these reservoirs has been comingled since 1960, with initial oil in place estimated to be some 14.5×10^6 m^3. Jurassic reservoirs in and around the Wapella and Red Jacket pools are estimated to have approximately 4.7×10^6 m^3 oil in place.

The Red Jacket pool began oil production in 1984. Most of the oil comes from well-developed tidal channel sandstones in Moosomin B. Reserves are estimated to be 2.6×10^6 m^3 in this pool. A single well in the Red Jacket pool (*i.e.* Tricent *et al.* Red Jacket 11-4-14-32W1) also produces from a sandstone in the lower Gravelbourg with an estimated 0.22×10^6 m^3 oil in place.

Several geological factors have led to the accumulation of hydrocarbons. The oil is postulated to have migrated from the underlying Souris Valley beds across the Mississippian subcrop via a fracture-fault system (van Delinder, 1984). The stratigraphic distribution and geochemical composition of this oil suggest that it came from a source in the Mississippian Madison Group (Brooks *et al.*, 1987).

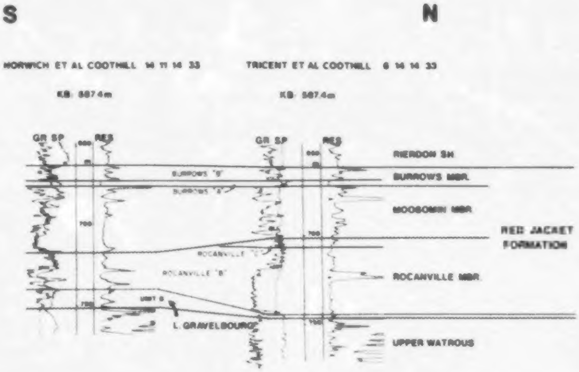


Figure 13: Cross-section showing erosion and infilling by Rocanville B strata of an area previously occupied by unit B of the lower Gravelbourg.

At present all known occurrences of Middle Jurassic oil in the southeast are found either beyond the zero limit of the lower Watrous red beds or where the lower Watrous is thin (Fig. 8). Thus the lower Watrous may have acted as a barrier to oil migration. However, very few wells have been drilled between the Wapella, Coothill and Red Jacket pools and the Mississippian oil pools 30 km to the south. The oil potential of the Jurassic in this area is relatively unknown.

Upper Watrous sediments separate the Mississippian subcrop from the overlying oil-prone Mesozoic section in the Wapella-Moosomin area, but show considerable evidence that they were not effective seals to upward oil migration, as they are commonly brecciated, oil stained, and display vertical fractures sealed by gypsum. Brecciation is attributed to exposure and weathering processes shortly after deposition, but the vertical fractures resemble those found in the underlying Souris Valley carbonates. It is likely that at least some fracturing occurred in response to the Laramide orogenic event of the Late Cretaceous to Tertiary. This fracturing appears to have facilitated migration of oil into overlying Jurassic and Cretaceous strata in the study area. A number of the localities showing structural highs are oil productive probably as a consequence of fracturing and doming related to the far-reaching effects of the orogeny.

It is probable that fracture systems, similar to those observed in the Mississippian and upper Watrous in the Wapella-Moosomin area, exist in the area to the south. If such fracture systems have breached both upper and lower Watrous strata then there is significant Jurassic and Cretaceous hydrocarbon potential in those areas.

Generally, the sandy peloidal bioclastic limestones and quartzose sandstones of the Lower Gravelbourg B unit are well-cemented with calcite. Nevertheless, oil is presently being produced from a weakly calcite-cemented sandstone in the Tricent *et al.* Red Jacket 11-4-14-32W1 well in the Red Jacket pool. The overlying *Modiolus* (?)-bearing mudstones of the unit provide the necessary seal for oil entrapment. Production from the lower Gravelbourg at Tricent *et al.* Moosomin 4-28-14-31W1 was also attempted but abandoned due to poor quality of the reservoir.

In the Rocanville B unit a sequence of predominantly near-shore marine shelf and relict beach sandstones form stratigraphic traps in the Wapella and Coothill pools. These sandstone bodies have an apparent sheet-like distribution and are stacked in places (Figs. 9 and 10). Some of the apparent stacking may be attributed to multi-stage salt solution effects, wherein anomalously thickened sands accumulated on actively subsiding solution/collapse surfaces. The anomalously thick sandstones at the base of Rocanville B at Tricent *et al.* Coothill 6-14-14-33W1 (Fig. 13) may have been formed by such a process. Alternatively, part of the thick sandstone section in this well may be a channel deposit at the base of Rocanville B.

Commonly calcite and dolomite cementation at or near the top of Rocanville B sandstones, or alternatively, overlying siltstones and dolomitic mudstones act as seals to these reservoirs. Although the traps appear to be largely stratigraphic there also appears to be a structural component to some of them. The multi-stage dissolution of the underlying Devonian Prairie Evaporite and the resulting collapse of overlying strata likely played a major role in the structural configuration. Both the Wapella pool and the area around Tricent *et al.* Coothill 6-14-14-33W1 show evidence of collapse structures and both are hydrocarbon producers (Fig. 14).

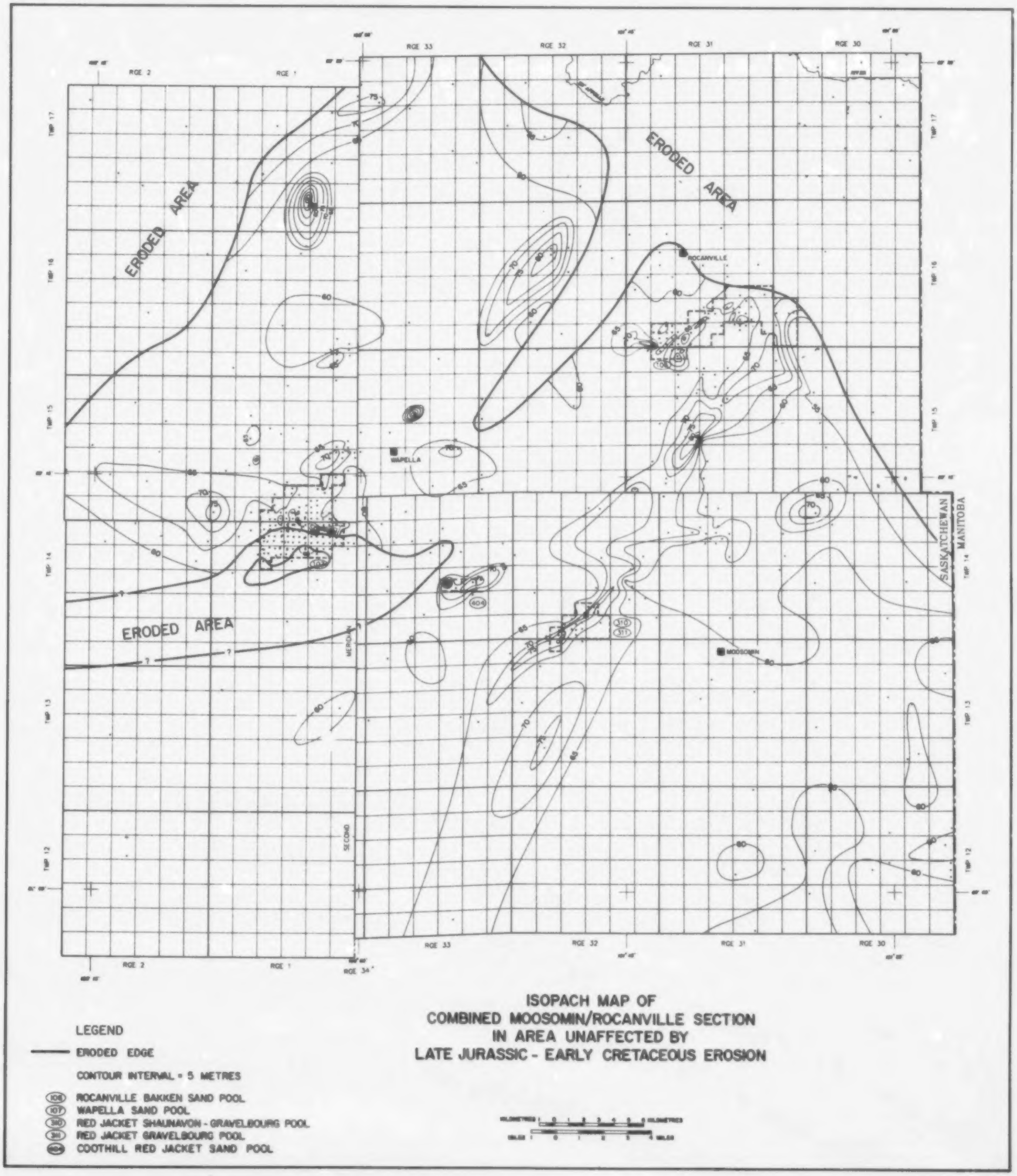


Figure 14: Isopach map of combined Moosomin/Rocanville section in area unaffected by Late Jurassic-Early Cretaceous erosion.

a

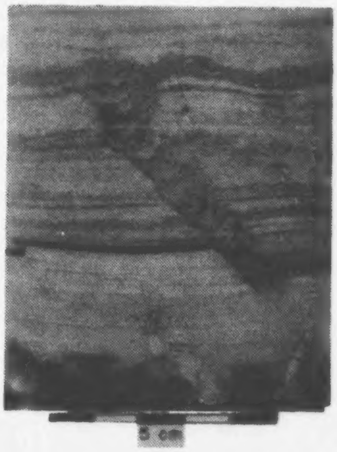
b

c

d

e

f



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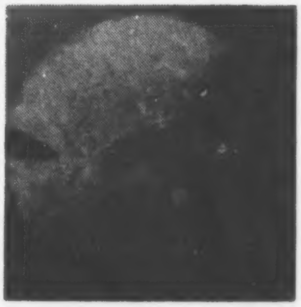

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Besides providing structures conducive to oil entrapment, collapse structures may also have provided pathways for the migration of oil up into traps. Targeting of salt dissolution/collapse structures may therefore help to locate hydrocarbon reservoirs in Middle Jurassic and/or Cretaceous reservoirs. Obviously, not all such sites would have hydrocarbon reservoirs despite their anomalously thick sandstones, *e.g.*, the Chevron Clayridge 8-10-17-1W2 well (Fig. 14), but such sandstones may be continuous and in a trapping configuration proximal to the barren site, perhaps updip to the north and east.

The Red Jacket pool and Moosomin locality lie within a northeasterly trend of anomalous thickening in the Wapella-Moosomin area (Fig. 14) which is interpreted to be a series of sinkholes developed by a salt-dissolution process. The near coincidence of the Red Jacket-Rocanville paleo-valley with this thickness trend strongly suggests that their geometries owe their form to a common deep-seated basement lineament. It is likely that structural movement related to the Nelson River gravity trend and the Birdtail-Waskada Axis played a role.

Moosomin A1 is found in a few wells of the Wapella pool (*e.g.* CNW Wapella 6-2-15-1W2). It is interpreted to be a sequence of fluvial or estuarine valley fill deposited in small channels or perhaps as crevasse splays, in the Wapella-North Wapella paleo-valley (Fig. 11). Prospects for reservoirs of this type would likely be greatest within, or on trend with, the Wapella-North Wapella and Red Jacket-Rocanville paleo-valleys (Fig. 11), along which these facies are most likely to occur.

In the Wapella pool, oil is also found in the overlying Moosomin A2 subunit which is interpreted to have been deposited as part of a barrier-island complex. Sandstones in this unit are commonly discontinuous and vertically stacked. Exploration for stratigraphic traps of this type is difficult, requiring numerous closely spaced wells to determine their complex distribution. Siltstone and mudstone interbeds often envelop discrete sandstone lenses in this unit creating the necessary seals for oil traps. In some wells, widespread dolomite and calcite cementation of sandstones at or near the top of Moosomin A effectively traps the oil (*e.g.* Saskoil Wapella 7-35-14-1W2).

Plate 1A: Oolitic and peloidal bioclastic limestone showing high-angle planar cross-stratification (approx. 25°) in the lower half of photo and small-scale graded bedding in the upper half. (Tricent *et al.* Moosomin 4-28-14-31 well (Lsd. 4-28 -14-31W1) at 654.0 m)

Plate 1B: Arrow indicates sharp and erosive contact with Rocanville A at base of high-angle planar cross-stratified sandstone in Rocanville B. Siltstone pebbles of same composition as A are incorporated in sandstone immediately above contact. (Pondex Pat Parkman 5-22-10-1 well (Lsd. 5-22-10-1W2) at approximately 964.0-967.0 m)

Plate 1C: Profile of mud cracks in variegated mudstone in Rocanville B of Rocanville Member. (Tide Water Wapella No 6-33 well (Lsd. 6-33-14-1W2) at approximately 745.2 m)

Plate 1D: Moosomin A1 sandstones and siltstones. Arrows on the left point to high-angle planar cross-bedding showing opposing flow directions (*i.e.* bipolar flow). The arrow on the right points to interlaminated siltstone and argillaceous sandstone. (CNW Wapella 6-2-15-1 well (Lsd. 6-2-15-1W2) between approximately 739.3-743.6 m)

Plate 1E: Burrow-mottled sandstone and mudstone in Moosomin B. (Tricent *et al.* Red Jacket 4-4-14-32 well (Lsd. 4-4-14-32W1) at approximately 582.4 m)

Plate 1F: *Diplocraterion* boring in a firm substrate (*Glossifungites* ichnofacies) near the top of Moosomin B. Chevron Whitewood 4-18-15-2 well (Lsd. 4-18-15-2W2) at approximately 723.9 m)

Plate 1G: *Diplocraterion* burrow in a shifting substrate (*Skolithos* ichnofacies) near top of Burrows B. It appears to be an escape burrow created by its occupant in response to rapid deposition of sand. (Tricent *et al.* Red Jacket 3-15-14-32 well (Lsd. 3-15-14-1W2) at approximately 654.3 m)

Plate 1H: Syneresis cracks in mudstone near top of Moosomin B. (Coho *et al.* Welwyn 8-16-15-31 well (Lsd. 8-16-15-31W1) at approximately 560.0 m)

Plate 1I: Arrow points to mud crack at top of Moosomin B which is infilled with sandy mudstone of overlying Burrows A. The contact is at 666.5 m in the Tricent *et al.* Red Jacket 12-4-14-32 well (Lsd. 12-4-14-32W1).

Plate 1J: Mudstone drapes (arrows) in tidal channel sandstone in Moosomin B. (Tricent *et al.* Red Jacket 2-4-14-32 well (Lsd. 2-4-14-32W1) at approximately 669.8 m).

Stratigraphic traps within Moosomin B represent the uppermost stratigraphic level of Jurassic hydrocarbon production in southeastern Saskatchewan. Most of the oil trapped in this unit occurs in tidal channel sandstones, as found in the Red Jacket pool, the North Wapella and Moosomin localities, and in the areas centred on the Coho *et al.* Welwyn 1-16-15-31W1 and Coho *et al.* Burrows 13-1-16-1W2 wells. The Red Jacket pool, the Moosomin locality and the area around Coho *et al.* Welwyn 1-16-15-31W1 align along a north-northeasterly trend (Fig. 1). These locations are within, or close to, the northeasterly trend of thickness anomalies shown in Figure 14, which are coincident with, or parallel to, the Red Jacket-Rocanville paleo-valley (Fig. 11). The North Wapella locality and the area centred on the Coho *et al.* Burrows 13-1-16-1W2 well also align along a north-northeasterly orientation which is coincident with, or parallel to, the Wapella-North Wapella paleo-valley. Again, areas of hydrocarbon accumulation appear to be closely associated with areas of salt dissolution/collapse structures.

Sandstone bodies of similar character and stratigraphic level to those found in Moosomin B are also oil reservoirs in an area centred on the Troy St. Lazare 1-21-16-29W2 well in southwestern Manitoba. This demonstrates the potential for hydrocarbon entrapment in Middle Jurassic sediments outside the area of this study.

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STRATIGRAPHY, PALEONTOLOGY, AND ASPECTS OF DIAGENESIS OF THE WHITEMUD FORMATION (MAASTRICHTIAN) OF ALBERTA AND SASKATCHEWAN

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ABSTRACT

The Maastrichtian Whitemud Formation crops out in river cuts and in relatively elevated areas in the plains of Western Canada from about the Edmonton area to the southeasternmost corner of Saskatchewan. Its area of deposition probably covered most of southern Alberta and Saskatchewan. In Alberta it is conformably overlain by the lacustrine Battle Formation which has yielded radiometric age of 65 Ma. East of the Cypress Hills it is overlain disconformably or paraconformably by younger sediments.

The Whitemud strata are composed of sandstones, mudstones and thin interbedded lignite seams. Grain size and facies analyses of these fining upward sequences indicate deposition of sediments in meandering stream environments with extensive lacustrine and swampy conditions.

Petrological differences between the Whitemud strata of the Alberta plains and of Saskatchewan suggest different sources of sediment. In Alberta the sands contain abundant volcanic rock fragments which produced montmorillonite by diagenetic alteration. In Saskatchewan, kaolinite is locally abundant as alteration product of feldspars and metamorphic rock fragments and probably also as detrital clay.

Vegetal remains and coprolites are the only fossils found in the Whitemud Formation. The typical Whitemud microfossil assemblage consists of megaspore genera *Azolla*, *Ghoshispora*, *Erlansonisporites*, by seed cuticle genera *Costatheca*, *Spermatites*, *Carpotheca*, and by palynomorphs of the *Wodehouseia spinata* Zone. This typical microfossil assemblage, which is consistently found in the Whitemud Formation from the Red Deer River Valley of Alberta to southern Saskatchewan, suggests synchronous deposition of the unit over the whole region.

In southern Saskatchewan, floodplain mudstones of the Whitemud Formation contain abundant vertebrate faecal remains preserved as siderite and Fe-hydroxides. The coprolites can be grouped into nine morphological classes on the basis of style of coiling, shape, size, and surface striations. Their morphology is consistent with deposition of the faeces by terrestrial reptiles. The excellent preservation of the fossils can be attributed to early diagenetic precipitation of $FeCO_3$ in loci where faecal matter was still bacterially fermenting.

Isotopic $\delta^{13}C$ and $\delta^{18}O$ ratios of coprolites are consistent with early diagenetic replacement of the faecal matter by siderite, followed by partial oxidation to Fe-hydroxide. ^{13}C ratios of siderite spherulites and concretions from the Whitemud Formation in the Red Deer River Valley and in the Cypress Hills also reflect the influence of biogenic carbon in a reducing, methanic, diagenetic environment.

INTRODUCTION

Following the withdrawal of the Campanian-Maastrichtian Bearpaw Sea (Caldwell, 1982), continental sediments were deposited over the plains of southwestern Canada. In response to pulses of uplift and periods of relative quiescence of the Columbian-Laramide Orogen to the west, meandering streams, coal swamps and extensive lakes characterized the late Maastrichtian and Paleocene landscape of southern Alberta and Saskatchewan. These sediments have been the subject of numerous investigations because of their economic potential (mainly coal and kaolin), of their vertebrate fossil remains and also because the Cretaceous/Tertiary (K/T) boundary occurs within them.

The Whitemud Formation, a sandy-clayey unit occurring approximately 50 m below the K/T boundary in the Red Deer

River Valley (Lerbekmo, 1985) can be easily recognized in the badlands of southern Alberta and in the Cypress Hills (Figs. 1 and 2) where it is overlain in chromatic contrast by the purplish brown Battle Formation. East of the Cypress Hills, where the Battle Formation is absent due to a disconformity (Furnival, 1946), the Whitemud Formation can still be recognized in river cuts and badlands in south-central Saskatchewan because of its strikingly white colour and kaolinitic nature.

In the mid 1960's the Whitemud and Battle Formations were thought to have been deposited under marine conditions. Later, it was shown that the Whitemud Formation has sedimentological characteristics typical of meandering streams and alluvial floodplains (Binda and Lerbekmo, 1973; Broughan, 1984). The overlying Battle Formation has been interpreted as a lacustrine deposit by Binda (1970 and 1991) and by Russell (1977). The terrestrial environment of the two units has been documented in a number of microfloral studies (Binda, 1968; Binda and Nambudiri, 1983 and 1988; Binda and Srivastava, 1968; Srivastava and Binda, 1969 and 1984).

In this paper we (1) outline the main stratigraphic, sedimentological, petrological, and mineralogical characteristics of the formation from the Red Deer River Valley of Alberta to south-central Saskatchewan, (2) provide a complete list of its fossils, (3) present new palynological data which allow bio- and chronostratigraphic correlation of the unit, and (4) present new isotopic data relevant to its diagenesis.

PHYSICAL STRATIGRAPHY AND SEDIMENTOLOGY

The type-section of the Whitemud Formation was selected by Kupsch (1956) in Dempster's kaolin quarry, in the eastern Cypress Hills (Figs. 1 and 3). There, as in a few other clay pits in the Cypress Hills which may be used as reference sections, the formation can be divided into three members: an upper, kaolinitic member, a middle, coaly member and a lower, sandy member. This subdivision has been accepted by Broughan (1984) who interpreted the facies sequence as evolving from channel and point-bar of meandering streams (lower member) to floodplain (middle and upper members). The coaly lenses of the middle member formed where local conditions of poor drainage and high water table developed on the floodplain, whereas the lithological assemblage of the upper member formed, according to Broughan, by overbank flow and deposition in small lakes.

In the Red Deer River Valley of the Alberta plains, the sedimentary rocks overlain by the Battle Formation and underlain by the Thomson coal seam have been assigned formational status and equated with the Whitemud Formation by Irish and Havard (1968). It is impossible to achieve a detailed lithostratigraphic correlation of these fluvial sediments covering an area of a few hundred thousand square kilometres, with limited outcrop, and gaps of hundreds of kilometres. As was pointed out by Binda and Lerbekmo (1973), the white-weathering, light grey sandstones and greenish grey mudstones which underlie the Battle Formation in the Alberta plains occupy the same stratigraphic position as the Whitemud Formation of the Cypress Hills. However, at places they are underlain by a coal seam (Fig. 3) and at several localities they grade downwards into darker grey sandstones.

It is tempting to correlate stratigraphically the whole of the Whitemud Formation of the Alberta plains with the upper member of the Whitemud Formation of the Cypress Hills using as a marker the coaly bed underlying both units (Fig. 3) but, since the continuity cannot be proven, the exercise would be futile.

Throughout the region, the Whitemud Formation displays a fining upwards sequence culminating in lacustrine sediments. Sedimentary structures are not always easy to detect in weathered outcrop; however, typical sequences from mud clasts at the base to cross-bedding to parallel bedding to rhizoconcretions, indicating sand bars migrating in channels and eventually being covered by flood plain sediments, can be seen at a few localities.

The interpretation of grain size distribution of Whitemud sandstones led to the same conclusions: "Conditions of fast-flowing water, such as can be encountered in river channels, are indicated mostly in the lower part of the section..... Energy conditions typical of floodplains and of quieter water, with deposition of large amounts of fine material, are more characteristic of the higher part of the section". (Binda and Lerbekmo, 1973, p. 72-73).

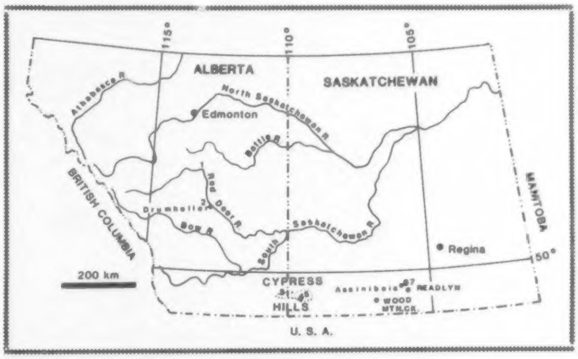

Figure 1: Map of southern Alberta and Saskatchewan showing location of lithostratigraphic columns 1-7 of Figure 3.

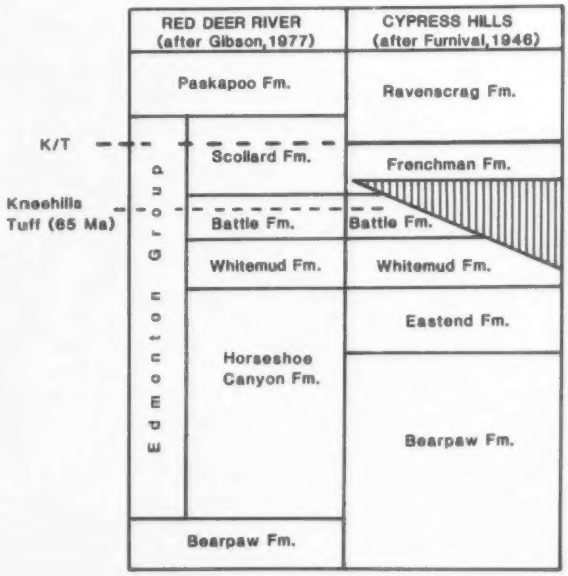

Figure 2: Correlation of Maastrichtian and Paleocene strata in southern Alberta and Saskatchewan.

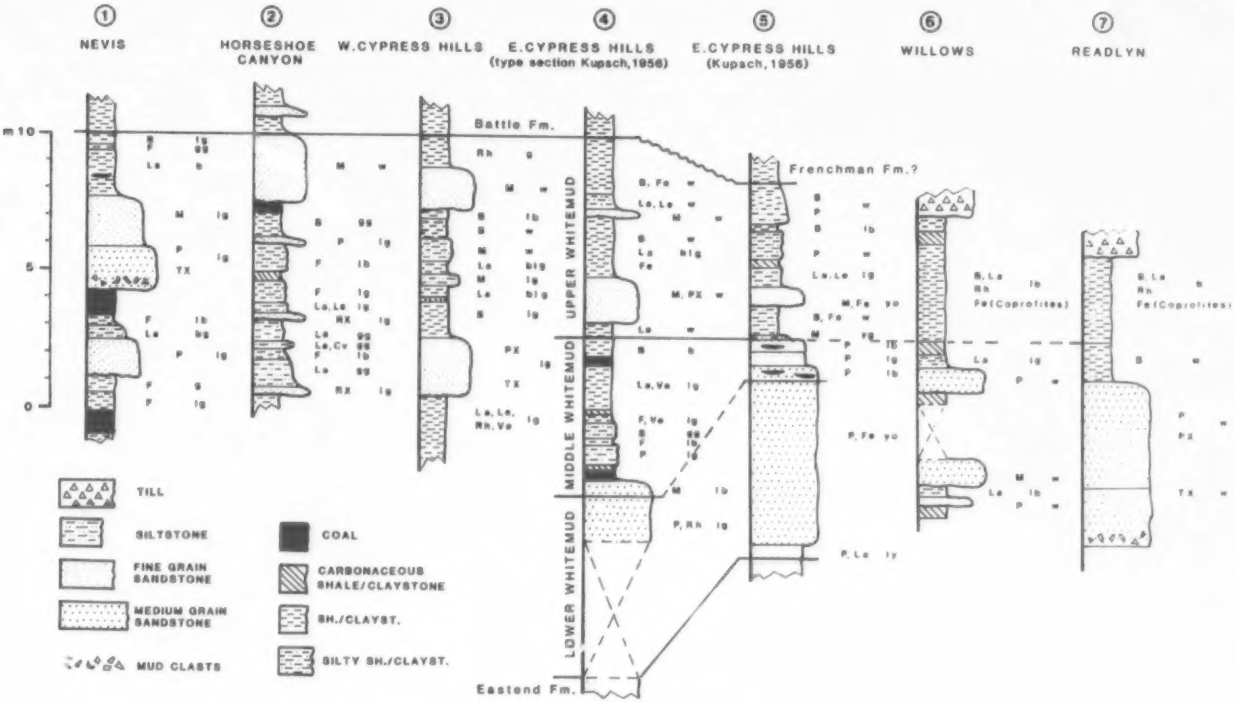

Figure 3: Lithostratigraphic columns of the Whitemud Formation in southern Alberta and Saskatchewan. Location of outcrops 1-7 is given in Figure 1.

Key to abbreviations:

STRUCTURES (left, in capital letters): B = blocky, Cv = convolute bedding, F = fissile, Fe = iron concretions, La = laminated, Le = lenticular bedding, M = massive, P = parallel stratification, PX = planar cross stratification, Rh = vertical root casts (rhizoconcretions), RX = ripple cross stratification, TX = trough cross stratification, Ve = vegetal fossil remains (leaves, stems).

COLOURS (right, in small-case): b = brown, bg = brownish grey, blg = bluish grey, g = grey, gg = greenish grey, lb = light brown, lg = light grey, ly = light yellow, w = white, y = yellow.

SANDSTONE PETROLOGY

The average composition of Whitemud sandstones from the Alberta plains, based on 200-point count of 23 thin sections is:

Quartz	27.0%
K-Feldspar	7.5%
Plagioclase	2.0%
Rock Fragments	39.0%
Accessories	2.5%
Matrix	22.0%

The major-components plot of the samples according to Travis (1955) classification is shown in figure 4. In this classification scheme, the sandstones can be termed lithic sandstones to rock-fragment sandstones.

One third to one half of the rock fragments are of volcanic origin and up to one quarter are chert. Metasediments and unidentified lithic clasts make up the remainder of the rock fragments. Quartz grains are mostly monocrystalline with straight to slightly undulose extinction. The Whitemud sandstones of the Alberta plains can thus also be termed "immature volcanic arenites".

CLAY MINERALOGY

Twenty-seven samples from the Whitemud Formation and nine samples from the Battle Formation of the Alberta plains and of the Cypress Hills were analyzed by X-ray diffraction by Binda (1970). The samples were prepared by settling the < 2 μm fraction on the glass slides and run, first untreated, then after treatment with ethylene glycol, and finally after oven-heating at 550° to 600° C, using a Philips Norelco diffractometer with nickel-filtered Cu K-alpha radiations at 35 KV and 15 MA. Examination of diffractograms in the region between $2\theta = 4$ and $2\theta = 20$ (22 Å to 4.4 Å) allowed the identification of the following groups of minerals on the basis of basal spacing.

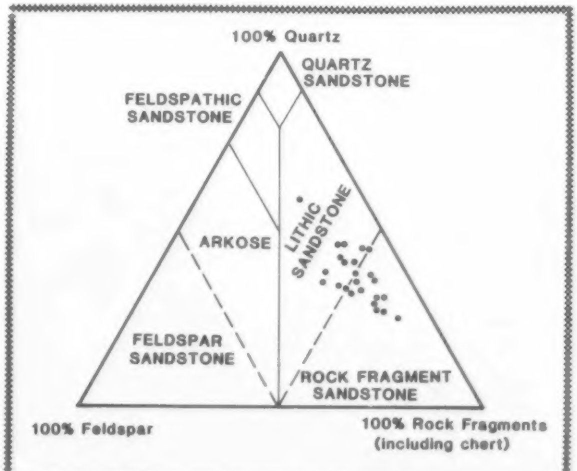

Figure 4: Major components compositional diagram of sandstones of the Whitemud Formation from the Alberta Plains.

12 Å group

Most untreated samples have a broad peak around 12 Å. The peak shifts to 16-17 upon glycolation and to 9-10 Å upon heating. This identifies smectite (Keller, 1962; Weaver, 1956; Warshaw and Roy, 1961). A shift to 17 Å upon glycolation (e.g. Fig. 5A, D, E) indicates more than 90 percent smectite, whereas a shift to smaller basal spacing (e.g. Fig. 5B) suggests a mixed layer clay containing 10 percent or more non-expandable layers. The 12 Å mineral of more than two-thirds of the analyzed samples is a mixed-layered clay containing 80-90 percent smectite. A shift to a 9-10 Å position upon heating suggests that most of the non-expandable layers are illite. However, if the shift is to a 9.3-9.7 Å position, the mixed-layer clay is likely to contain some chlorite (Weaver, 1956). Thus, sample B of figure 5 should contain a mixed-layer clay composed of approximately 90 percent expandable smectite (16.8 Å peak of glycolated sample) and almost 10 percent illite with traces of chlorite (9.71 Å peak of heated sample).

7 Å group

Peaks in the 7 Å basal spacing region, which are not affected by glycolation and disappear upon heating the sample at 550°-600° C are attributed to kaolinite. Halloysite would show broader peaks than the ones of Figure 5 and dickite has a dehydroxilation temperature of 650° C. The sharp peaks of the Cypress Hills samples (Fig. 5C and D) are indicative of well crystallized kaolinite. Small amounts of chlorite are difficult to detect in the presence of kaolinite and smectite. Small peaks in the 13-14 Å regions in diffractograms of heated samples (e.g. Fig. 5B) may be caused by small amounts of chlorite. However, the 7 Å peaks of the Whitemud samples are entirely attributable to kaolinite rather than to second order reflections of chlorite as can be ascertained from higher order kaolinite peaks in the 1.48 - 1.49 Å region, as well as from examination of thin sections.

9 - 10 Å group

Peaks in the 9 - 10 Å region indicate minerals of the illite (hydrous mica) - muscovite group. The peaks are affected neither by glycolation nor by heating. Sharp and narrow peaks such as the ones of samples C and D of figure 5 can be interpreted as due to fine grained muscovite. On the other hand, broad, low arched or poorly defined peaks such as the 9.82 Å peak of glycolated sample B of figure 5 may be attributed to illite.

Relative abundance of clay minerals

Estimates of relative abundances based on peak areas show that the dominant clay mineral in samples of the Whitemud Formation from the Alberta plains is a mixed-layer clay containing approximately 90 percent smectite. Kaolinite is commonly present in lesser amounts and illite is occasionally present in minor to trace amounts. Some samples from the Cypress Hills are also rich in smectite; however, major amounts of kaolinite and hydrous mica occur in all the samples from the Cypress Hills. Kaolinite is quarried commercially from the Whitemud Formation at several localities in the Cypress Hills and in south-central Saskatchewan, but not in the Alberta plains.

This regional difference in clay mineralogy can possibly be explained as due to a combination of detrital inheritance and neoformation. The sediments of the Alberta plains are rich in volcanic rock fragments whereas those of the Cypress Hills are richer in metamorphic fragments. Mellon (1960) showed that in the Lower Cretaceous rocks of Alberta, kaolinite is the authigenic

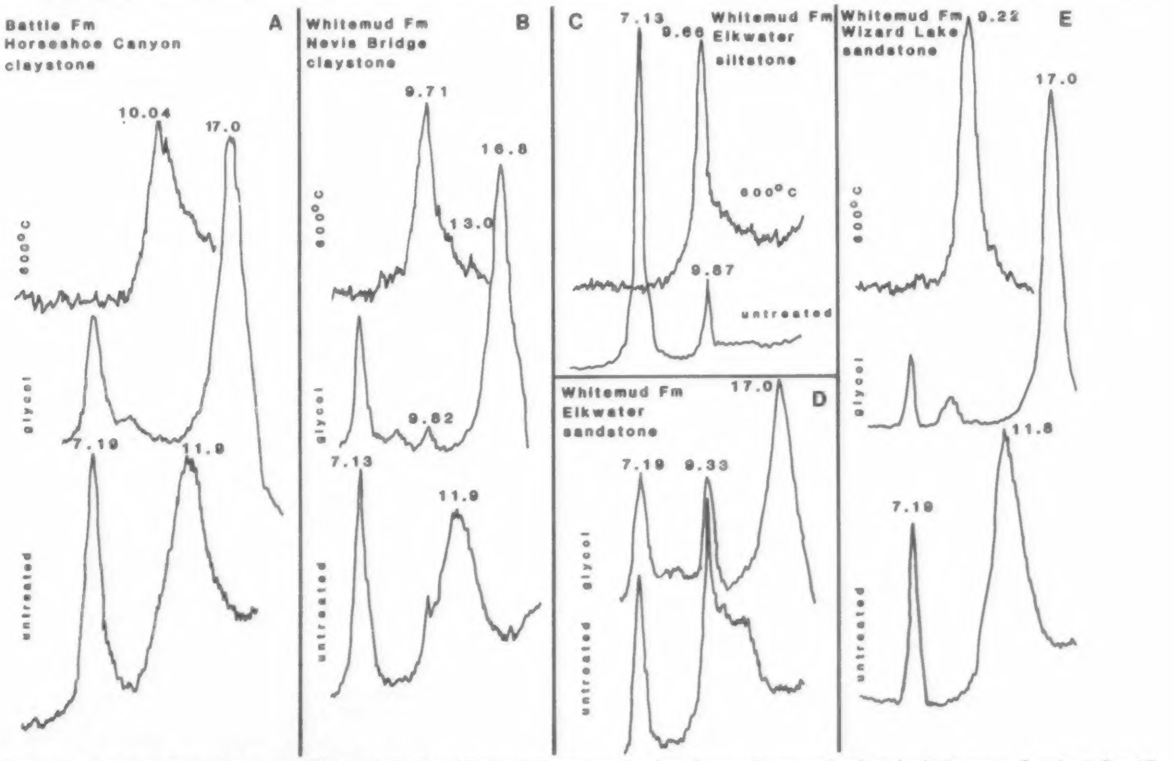

Figure 5: Partial x-ray diffractograms of clays of the Whitemud (B - E) and Battle (A) Formations. Numbers above peaks express basal spacing in Angstrom. Samples A, B and E are from the Alberta plains, samples C and D from the Cypress Hills.

clay mineral of the sands low in volcanic detritus, whereas smectite is the common authigenic clay of the sands rich in volcanic fragments. It can also be speculated that, besides the authigenic components, a detrital smectitic clay was released from the source area that provided the volcanic rock fragments, and kaolinite together with illite and/or sericite from the source that provided metamorphic rock fragments.

PROVENANCE

Sandstone petrology and clay mineralogy suggest different provenances for the Whitemud sediments of the Alberta plains and of the Cypress Hills-Saskatchewan. The most likely source for the sediments of the Alberta plains was to the west, where mountain-building had been in progress since early Mesozoic time. The source area of the Whitemud Formation of the Cypress Hills is less easy to identify. The abundance of metamorphic clasts would suggest a provenance from the Canadian Shield to the north or northeast. However, Byers (1969) argued for a more proximal source to the southwest, in Montana, to account for the angularity of the grains and for the metamorphic-chert-volcanic clast assemblage.

PALEONTOLOGY

Only vegetal fossil remains and trace fossils have been found in the Whitemud Formation: vertebrate and invertebrate fossils appear to be conspicuously absent from this unit. A list of plant fossils of the Whitemud Formation is given in Tables 1 and 2. Some typical fossils are illustrated in Figures 6 and 7.

PLANT MACROFOSSILS

From Wood Mountain, southern Saskatchewan (Fig. 1) Berry (1935) reported twelve species of leaf impressions, three stems and twigs of conifers and angiosperms, and one fruit which he considered related to modern Cercidiphyllaceae. Bell (1949) revised Berry's taxonomic work and suggested for the leaves of the Whitemud Formation affinities with Ginkgoaceae, Cupressaceae, Nymphaeaceae, etc. (Table 1). However, caution must be exercised in accepting these affinities, since much of the early paleobotanical work on leaf impressions in North America has not withstood the test of time.

Nambudiri and Binda (1989) reported from the Assiniboia area a petrified dicotyledonous fruit, *Limalbocarpon readlynensis* resembling extant fruits of *Phyllanthus emblica* of Euphorbiaceae, together with other fruits of unknown affinity, assigned to the genus *Carpolithes* Brongniart 1828.

SEED CUTICLES

Several species of seed cuticles belonging to three genera, *Spermatites*, *Costatheca* and *Carpotheca* were described from the Whitemud Formation by Binda (1968, 1970) and by Binda and Nambudiri (1983). While species of *Spermatites* and *Costatheca* are abundant in samples from the Red Deer River Valley, *Carpotheca* species are more common in the sections located to the south and east (Binda and Nambudiri, 1983). Species of *Spermatites* and *Costatheca* have also been reported from other sites in the western Canadian Plains and the Alberta foothills (Gunther and Hills, 1972; Kutluk and Hills, 1984; Speelman and Hills, 1980). These seed cuticles were regarded as bryophyte remains by Miner (1935) and as insect eggs by Tiffney (1983). Binda (1968) showed similarities between *Spermatites* and seed cuticles of extant *Juncus*. Binda and Nambudiri (1983) discussed similarities between *Costatheca* and *Butomus umbellatus* and between *Carpotheca coronata* and *Typha latifolia* and *T. angustifolia*.

MEGASPORES

Megaspores that are commonly encountered in the Whitemud Formation in Alberta and Saskatchewan are *Azolla* and *Ghoshispora* (Table 1). Along with these genera can be found megaspores of *Erlansonisporites* and *Minerisporites*. Snead (1969) reported several species of Azolla from the Maastrichtian strata in central Alberta. Megaspores of this lacustrine, leptosporangiate fern genus, comparable to *Azolla distincta*, *A. filosa* and *A. barbata* (Snead, 1969) occur in the Whitemud Formation in Alberta and Saskatchewan.

Four species of *Ghoshispora* S.K. Srivastava (1966) are found in the Whitemud Formation. Several species of these acrolamellate megaspores, previously described by Srivastava and Binda (1969) under the form genus Balmeisporites Cookson and Dettmann 1958 were transferred by Srivastava (1978) into the genus *Ghoshispora*. The megaspores of *Ghoshispora* were produced by heterosporous aquatic ferns (Kondinskaya, 1966) probably related to Azollaceae. Although species of *Ghoshispora* have been reported from saline, brackish and paludal environments (Kutluk and Hills, 1984), the forms of the Whitemud Formation occurring with megaspores of Azolla, were undoubtedly deposited in freshwater environment.

Table 1: List of leaf impressions, stems and twigs, fruits, seeds, seed cuticles, and megaspores reported from the Whitemud Formation of Alberta and Saskatchewan and their probable affinities.

Taxon	Probable Affinities
LEAF IMPRESSIONS	
Ginkgoites adiantoides (Unger) Bell 1949	Ginkgoaceae*
Sequoiites artus Bell 1949	Cupressaceae*
Thuites interruptus (Newberry) Bell 1949	Cupressaceae*
Trochodendroides arctica (Heer) Bell 1949	Cercidiphyllaceae*
Menispermites belli Berry 1935	Menispermaceae*
Nelumbites protoluteus (Berry) Bell 1949	Nymphaeaceae*
Nymphaeites angulatus (Newberry) Bell 1949	Nymphaeaceae*
Nymphaeites striatus (Berry) Bell 1949	Nymphaeaceae*
Celastrinites insignis (Heer) emend. (Hollick) Bell 1949	Celastraceae*
Vitis stantoni (Knowlton) Brown 1939	Vitaceae*
Viburnum simile Knowlton 1930	Caprifoliaceae*
Pistia corrugata Lesquereux 1878	Araceae*
STEMS AND TWIGS	
Sequoiites artus Bell 1949	Cupressaceae*
Thuites interruptus (Newberry) Bell 1949	Cupressaceae*
Nymphaeites angulatus (Newberry) Bell 1949	Nymphaeaceae*
FRUITS AND SEEDS	
Jenkinsella arctica (Heer) Bell 1949	Cercidiphyllaceae*
Limalbocarpon readlynensis Nambudiri & Binda 1989	Euphorbiaceae
Carpolithes Brongniart 1828 Type A Nambudiri & Binda 1989	Unknown
Carpolithes Brongniart 1828 Type B Nambudiri & Binda 1989	Unknown
Carpolithes Brongniart 1828 Type C Nambudiri & Binda 1989	Unknown
Carpoilthes Brongniart 1828 Type D Nambudiri & Binda 1989	Unknown
SEED CUTICLES	
Genus *Spermatites* Miner 1935 emend. Binda & Nambudiri 1983	Juncaceae
Spermatites nanus Miner 1935	
Spermatites ellipticus Miner 1935 Subsp. *minor* Binda 1968	
Spermatites minimus Binda 1968	
Spermatites piperiformis Binda 1968	
Spermatites saskatchewanicus Binda & Nambudiri 1983	
Spermatites pinguis Binda & Nambudiri 1983	
Spermatites robustus Binda & Nambudiri 1983	
Genus *Costatheca* Hall 1967 emend. Binda & Nambudiri 1983	Butomaceae
Costatheca striata (Dijkstra 1949) Hall 1967 emend. Binda & Nambudiri 1983	
Costatheca inflata Binda & Nambudiri 1983	
Costatheca minerii Binda & Nambudiri 1983	
Costatheca exilis (Vangerow 1954) Hall 1967 emend. Binda & Nambudiri 1983	
Costatheca lata (Vangerow 1954) Hall 1967 emend. Binda & Nambudiri 1983	
Costatheca pusilla Binda & Nambudiri 1983	
Genus *Carpotheca* Binda & Nambudiri 1983	Typhaceae
Carpotheca elegans Binda & Nambudiri 1983	
Carpotheca coronata Binda & Nambudiri 1983	
Carpotheca parva Binda & Nambudiri 1983	
Carpotheca falcata Binda & Nambudiri 1983	
MEGASPORES	
Genus *Ghoshispora* S.K.Srivastava 1966	Azollaceae
Ghoshispora canadensis (S.K.Srivastava & Binda 1969) S.K.Srivastava 1978	
Ghoshispora bella (Kondinskaya 1966) S.K.Srivastava 1978	
Ghoshispora dettmannii (S.K.Srivastava & Binda 1969) S.K.Srivastava 1978	
Ghoshispora kondinskayae (S.K.Srivastava & Binda 1969) S.K.Srivastava 1978	
Genus *Erlansonisporites* Potonie 1956	Selaginellaceae
Erlansonisporites whitemudensis Kovach & Batten 1989	
Erlansonisporites robustus Binda & Nambudiri 1988	
Genus *Azolla* Lamarck	Azollaceae
Azolla filosa Snead 1969	
Azolla distincta Snead 1969	
Azolla barbata Snead 1969	
Genus *Minerisporites* Potonie 1956	Isoetaceae
Minerisporites mirabilis (Miner) Potonie 1956	

*Taxonomic assignment follows Bell (1949)

Binda and Nambudiri (1988) reported two new species of *Erlansonisporites*: *E. singhii* and *E. robustus* from the Whitemud strata in Alberta. *E. singhii* has since been transferred to *E. whitemudensis* Kovach and Batten 1989. *Erlansonisporites* is closely similar to megaspores of the heterosporous lycopod genus, *Selaginella* (Binda and Nambudiri, 1988). Megaspores of *Minerisporites mirabilis* (Miner) Potonié 1956 have been compared to those in the extant genus *Isoetes* (Kovach and Dilcher, 1988).

POLLEN AND SPORES

In his study of the Edmonton Group in the Red Deer River Valley, Srivastava (1970) found the Whitemud Formation to be a pollen-depauperate interval and thus called it the *Azolla-Balmeisporites* (now *Ghoshispora*)-bearing Interval. It is now clear that at least one sample assigned by Srivastava to the *Wodehouseia spinata* Zone is in fact well within the Whitemud Formation (3 m below the Battle/Whitemud contact). The list of pollen and spores contained in that sample (Scollard 18-13) and in six samples of the (upper ?) Whitemud Formation at Willows, Saskatchewan (Fig. 1), and of their relative abundances, is given in Table 2. Fifteen of the species in the Willows sections do not occur in the Scollard locality. On the other hand, seven species are restricted to the Scollard locality. Of the species that occur in both localities, two that are particularly useful in regional correlations are *Ulmipollenites planeriformis* and *Wodehouseia spinata*.

Nambudiri and Binda (unpublished) also found the following assemblage in a claystone 0.5 m below the Battle/Whitemud contact at Eastend in the Cypress Hills (Fig. 1): *Aquilapollenites delicatus* var. *collaris*, *A. reticulatus*, *A. reductus*, *Kurtzipites trispissatus*, *Wodehouseia spinata*, *Mancicorpus senonicus*, *Liliacidites mirus*, *Tricolpites microreticulatus*, *Polyporina sp.*, *Laevigatosporites ovatus* and *Taxodiaceaepollenites hiatus*. This assemblage is also comparable to the *Wodehouseia spinata* Zone (Zone VIII) of Srivastava (1970).

Thus, although the lower part of the Whitemud Formation, particularly in Alberta, must still be considered part of the pollen-depauperate *Azolla-Ghoshispora*-bearing Interval, probably because of poor preservation, we can now show the existence of a *Wodehouseia spinata* assemblage in the upper part of the formation at least at three widely separated localities.

TRACE FOSSILS (COPROLITES)

Vertebrate faecal remains preserved as siderite and Fe-hydroxides occur in great abundance in kaolinitic claystone of the Maastrichtian Whitemud Formation of southern Saskatchewan. Broughton *et al.* (1978) described several types of coprolites and attributed them to sturgeons (Acipenseridae) and bowfins (Amiidae). Broughton (1981) reported the additional find of up to 1 m long fossilized faecal matter and interpreted it as internal casts of the gut of teleostean fish or aquatic tetrapods.

Recently Schmitz and Binda (1991) examined several hundred specimens and grouped them into nine groups according to morphological features such as coiling, tapering, elongation and fine details of the surface. The morphology of the coprolites from the Whitemud Formation of southern Saskatchewan is the best evidence of their origin as excreta. Coiling, tapering, and fine striations cannot be explained by inorganic processes. Coiling is best explained by falling to the ground of a plastic faecal rope. Tapering at the extremities is obviously attributable to sphincter pinching. Longitudinal grooves and striations must be interpreted as the reflection of the internal morphology of the intestine and the sphincter-ring. They may be the result of incision of the faecal rope by anal papillae upon extrusion.

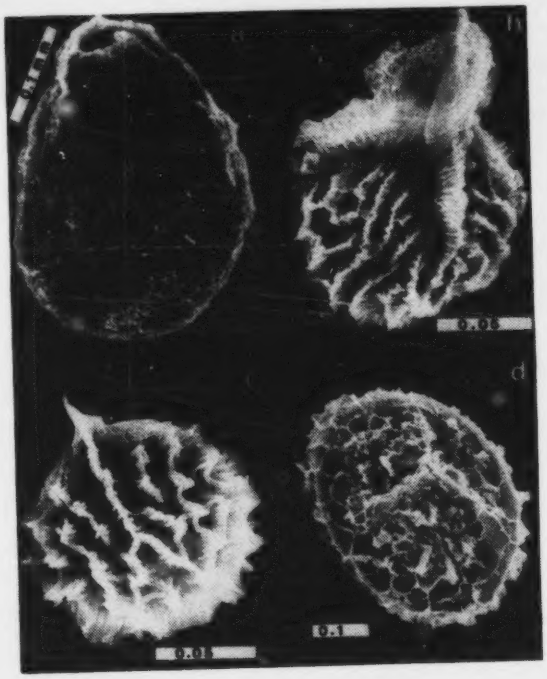

Figure 6: Typical megaspores of the Whitemud Formation.
a: *Azolla sp.*
b: *Ghoshispora kondinskayae*
c: *Ghoshispora bella*
d: *Erlansonisporites whitemudensis*

From comparison with present-day faeces it can be speculated that some rough-textured coprolites may have been produced by herbivorous reptiles and some smooth-textured ones by carnivorous. Only small rope-like, slightly sigmoidal coiled coprolites bear some resemblance to teleostean fish excreta. Unfortunately no trace of undigested content can be found in the coprolites, and our speculations are based only on the external morphology of the fossils which was subjected to modification during desiccation and diagenesis.

Additional evidence that the coprolites of the Whitemud Formation are faeces of land vertebrates and not of fish can be found in the mode of occurrence of the fossils. The coprolites occur only in the floodplain facies and have not been found in the cross-bedded point-bar facies. As noted by Nambudiri and Binda (1989), a large proportion of the coprolites are sun-cracked and have a flattened lower surface with petrified twigs adhering to it. This indicates subaerial deposition on a surface covered with vegetal litter, a condition that is compatible with a vegetated floodplain bordering a meandering stream.

The petrified faeces range from less than 1 cm to 7 cm in diameter. As was noted by Schmitz and Binda (1991), since upon drying, vertebrate faeces shrink to half of their original diameter, some of the Whitemud coprolites must have been excreted by animals capable of producing faeces of 14 cm in diameter. Thus, in the absence of large mammals during the Maastrichtian, the producers of the largest coprolites must have been large land reptiles.

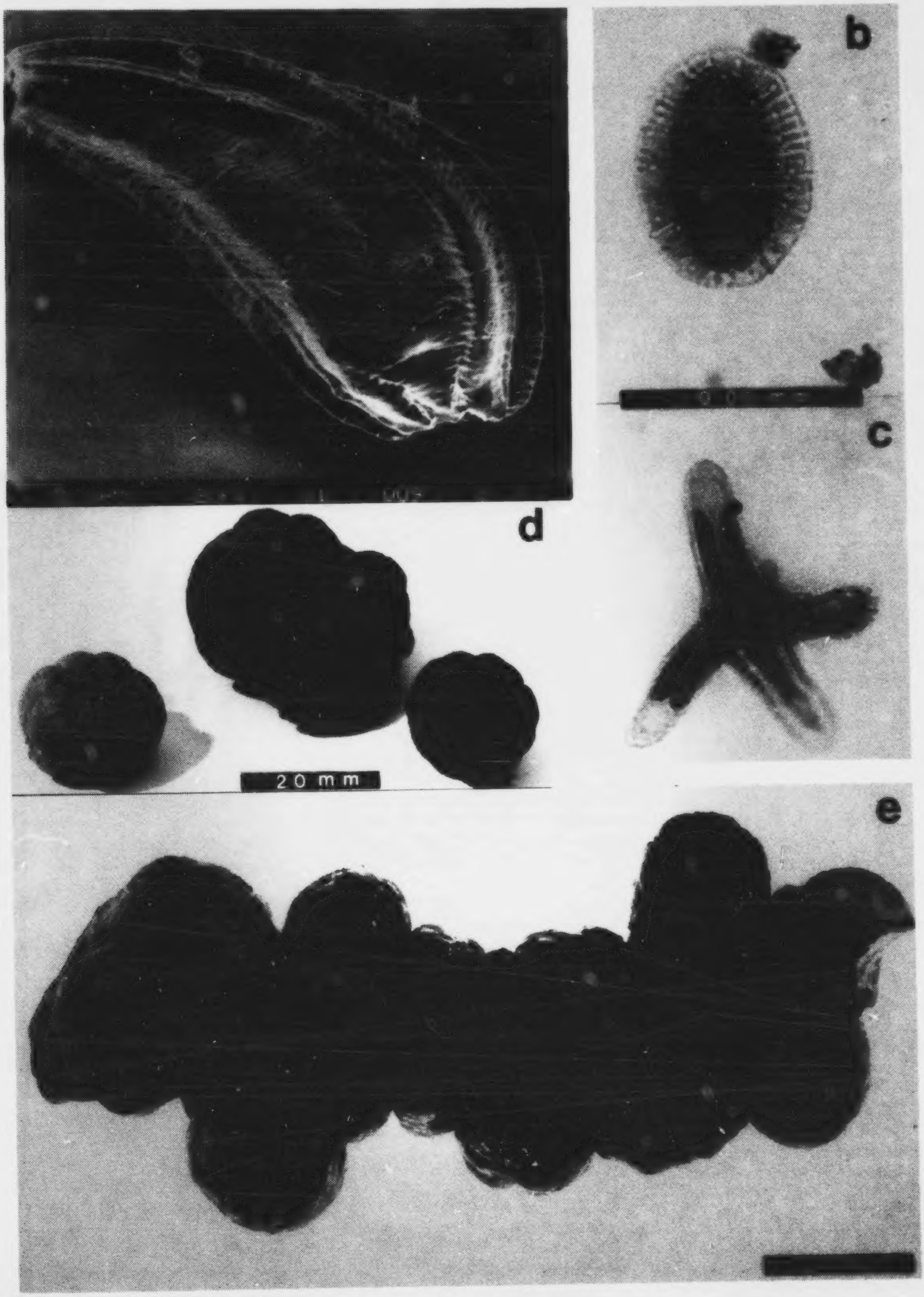

Figure 7: Typical fossils of the Whitemud Formation
a: *Costatheca minerii*; b: *Wodehouseia spinata*; c: *Aquilapollenites delicatus*; d: *Limalbocarpon readlynensis*; e: Coprolite

Table 2: Abundance, frequency, and probable affinities of pollen and spores at Scollard locality 18, Alberta (Srivastava, 1970) and in six samples from the Willows section of southern Saskatchewan.

Taxon	Scollard Alberta 18/13	Willows Section Saskatchewan						Probable Affinities
		1	2	3	4	A	D	
Albertipollenites rosalindiii S.K.Srivastava 1969a					R			Uncertain angiosperm (dicot)
Alisporites grandis (Cookson) Dettmann 1963	C	C	C		A	A	C	Bisaccate conifer
Alnipollenites finitimus (Chlonova) S.K.Srivastava 1968a	C	C			R			*Alnus*
Alnipollenites specipites Wodehouse 1933		C						*Alnus*
Aquilapollenites amicus S.K.Srivastava 1968b		A	R					? Lorathaceae
Aquilapollenites ascriptivus S.K.Srivastava 1969b		C						? Loranthaceae
Aquilapollenites aucellatus S.K.Srivastava 1969b		R						? Loranthaceae
Aquilapollenites dolium (Samoilovitch) S.K.Srivastava 1968b	C							? Loranthaceae
Aquilapollenites pumilis S.K.Srivastava 1969b	C							? Loranthaceae
Araucariacites australis Cookson 1947	C			R	C	C	R	Arauracaria
Carpinipites ancipites (Wodehouse) S.K.Srivastava 1966		C	C		R			*Carpinus*
Cicatricosisporites cuneiformis Pocock 1965 (RW)			C		R			*Mohria* (pteridophyte)
Classopollis classoides Pflug 1953 (RW)	R				R			Reworked gymnosperm)
Cranwellia striata (Couper) S.K.Srivastava 1966			R					Loranthaceae
Cycadopites carpentieri (Delcourt & Sprumont) Singh 1964	C				R	C		Cycad
Erdtmanipollis cretacea (Stanley) S.K.Srivastava 1972a	C							Buxaceae (*Pachysandra*)
Ghoshispora scollardiana S.K.Srivastava 1966	C	R						Azolla like aquatic fern
Deltoidospora minor (Couper) Pocock 1970					A	C		Cyatheaceae
Ericipites longisulcatus Wodehouse 1933						C	C	Ericaceae
Gleicheniidites senonicus Ross 1949 (RW)					C			Gleicheniaceae
Hazaria sheopiarii S.K.Srivastava 1971			R					? Polypodiaceae
Kurtzipites trispissatus Anderson 1960	C	A	C			A		Uncertain angiosperm (dicot)
Laevigatosporites discordatus Pflug 1953		A	A		R			Polypodiaceae
Laevigatosporites ovatus Wilson & Webster 1946	C	A	A		R			Polypodiaceae
Laricoidites gigantus Brenner 1963	C				C	C		*Larix*
Liliacidites mirus S.K.Srivastava 1969c			C		C			Yucca like plant
Liliacidites variegatus Couper 1953	C		C		C			Liliaceae
Lusatisporis dettmannae (Drugg) S.K.Srivastava 1972b	R	R						*Selaginella*
Mancicorpus senonicus Mtchedlishvili 1961		A	C					? Loranthaceae
Nyssapollenites parvus S.K.Srivastava 1969c	R							*Nyssa*
Osmundacidites wellmanii Couper 1953	C				R	R		Osmundaceae
Pandaniidites radicus Leffingwell 1971	C	R						*Pandanus*
Polyporina cribraria S.K.Srivastava 1969c	C	R	A	R	R			Chenopodiaceae
Quercoidites sternbergii S.K.Srivastava 1966	C	R				R		*Quercus*
Rhoipites crassus Stanley 1965	C							*Rhus*
Rutihesperipites trochuensis (S.K.Srivastava) S.K.Srivastava 1977	C		R					Aceraceae
Sigmopollis carbonis (Newman) S.K.Srivastava 1984	C	C	C					Algal zygote
Sigmopollis hispidus Hedlund 1965	R	R		R				Algal zygote
Taxodiaceaepollenites hiatus Kremp 1949	C			C		C		*Taxodium*
Tricolpites microreticulatus Belsky, Boltenhagen & Potonie 1965	C							*Gunnera*
Triporoletes stellatus S.K.Srivastava 1972b	C							Bryophyte
Ulmipollenites planeriformis (Anderson) S.K.Srivastava 1969c	A	A						*Ulmus*
Vitreisporites pallidus (Reissinger) Nilsson 1958 (RW)	R				C			Bisaccate conifer
Wodehouseia spinata Stanley 1961	A	A	C					Uncertain angiosperm

R = rare: 1-2 specimens; C = common: 3-5 specimens;
A = abundant: 6+ specimens; (RW) = reworked taxa;
*: significant for regional correlation

AGE AND CORRELATION

Isotopic dates, palynology, and magnetostratigraphy concur in assigning a late Maastrichtian age to the Whitemud Formation. K/Ar dates on sanidine and biotite in bentonitic tuff of the Battle Formation obtained by Folinsbee *et al.* (1965) are in the 64-67 Ma range at Strawberry Creek, near Edmonton, and 65 Ma in the western Cypress Hills (Fig. 1). The dates are considered analytically reliable by the authors who do not give upper and lower limits but suggest that the age of the bentonites is 65-66 Ma. The Whitemud Formation lies 5 to 8 m below the bentonites and there are no hiatuses in the intervening strata. Therefore, at the 35 m per million years sedimentation rate calculated by Lerbekmo (1985) for these sediments, the base of the Whitemud Formation should be no more than 0.8 Ma older than the bentonitic tuff and the top could be as little as 0.14 Ma older than the tuff. Comparable values can be obtained using the 64.3 ± 1.2 Ma multi-method average age of the K/T boundary (Baadsgaard *et al.*, 1988) which is 50 m stratigraphically above the Whitemud Formation in the Red Deer River Valley.

Although apparently devoid of macrofossils, the Battle-Whitemud interval marks an important boundary in dinosaur biostratigraphy in the region. Strata overlying the Battle Formation contain the typical *Triceratops* fauna of the Lancian Stage, whereas strata underlying the Whitemud Formation contain duckbill dinosaur fossils characteristic of the Edmontonian (Fox Hill) Stage (Sternberg, 1947; Russell and Chamney, 1967).

The late Maastrichtian age of the Whitemud Formation is confirmed by magnetostratigraphy. In the Red Deer River Valley, where sedimentation was continuous from Maastrichtian to Paleocene, the unit corresponds with magnetic polarity 30 Normal (Lerbekmo, 1985) which plots at approximately 66 Ma in the time scale given in Boggs (1987, p. 568).

In the Red Deer River Valley of Alberta, the Edmonton Group was subdivided by Srivastava (1970) into nine pollen assemblage zones and three sub-zones on the basis of angiosperm pollen alone. As stated in the paleontology section of this paper, the whole of the Whitemud Formation was then considered a pollen-depauperate interval, but it is now clear that at least the uppermost three meters of the formation in the Red Deer River Valley must be included in the *Whodehouseia spinata* Zone which extends up to the K/T boundary.

Species of the genus *Wodehouseia*, an angiosperm pollen of uncertain affinity, are excellent index fossils in the Upper Cretaceous stratigraphy of the Western Canadian Plains. Lerbekmo and Coulter (1985) correlated the Campanian - Maastrichtian boundary with the appearance of *W. edmonticola*. Similarly Lerbekmo (1985) correlated the appearance of *W. fimbriata* with the top of paleomagnetic reverse polarity zone 29 at the Maastrichtian-Paleocene boundary. *W. spinata* has been recorded throughout Lancian strata in western North America becoming extinct at the Cretaceous-Tertiary boundary (Stanley, 1965; Srivastava, 1970; Nichols *et al.*, 1986).

As much as there is consensus on the age of the Whitemud Formation, its precise chronostratigraphic correlation from the Red Deer River Valley to the Cypress Hills (Fig. 1) is controversial. Binda (1970 and 1991) pointed out that the identical opaline microfossil assemblage of megaspores, Chrysomonadinae, and silicosponge spiculae in the Battle Formation, as well as the identical megaspore seed cuticle assemblage in the Whitemud Formation at the two localities, make a strong argument in favour of synchronous deposition of the Whitemud-Battle package throughout the region. The presence of the Kneehills Tuff bed near the top of the Battle Formation in the Red Deer River Valley and in the western Cypress Hills (Ritchie, 1969; Binda, 1969) strengthens the argument considerably.

However, Lerbekmo's (1985) attempt at correlating the magnetic polarity zones of the terminal Cretaceous between Red Deer River Valley and Cypress Hills, revealed a telescoping between the 29 and the 31 Normal polarity zones in the Cypress Hills, in correspondence with the Frenchman-Battle Formation erosional gap. The impasse was resolved by assigning the Whitemud-Battle of the Cypress Hills to polarity zone 31 Normal. Thus, Lerbekmo concluded that the package is one half million years older in the Cypress Hills than in the Red Deer River Valley.

Comparison of pollen and spore assemblages from the Scollard locality 18, Red Deer River Valley (Srivastava, 1970) and from Willows, southern Saskatchewan (Table 2), clearly indicates biostratigraphic correlation of the Whitemud Formation of Alberta and Saskatchewan (Fig. 8). The correlation is strengthened by the occurrence of key elements of the same assemblage near the top of the Whitemud Formation at Eastend, in the Cypress Hills (Fig. 1).

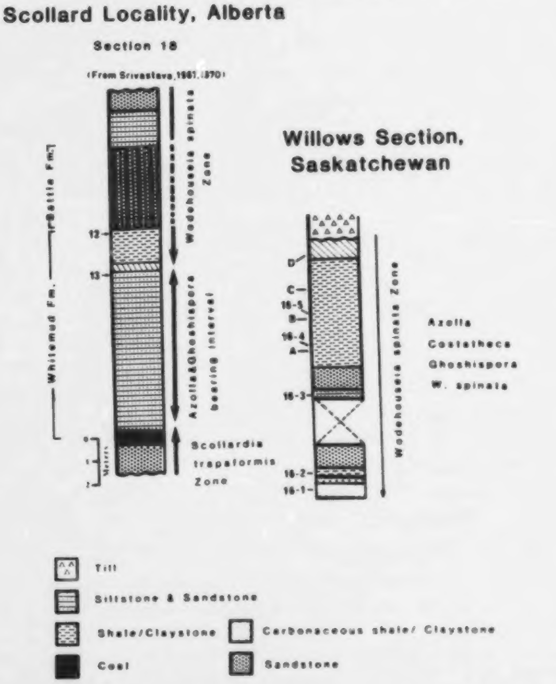

Figure 8: Stratigraphic relationship between the Scollard locality section in Alberta and the Willows section in southern Saskatchewan showing correlation between palynomorph assemblages.

Although more work is needed to identify with certainty the earliest occurrence of *Wodehouseia spinata* in Saskatchewan, the earliest known occurrence of the species in both provinces is in the Whitemud Formation. This reinforces the evidence in favour of synchroneity of the formation in the region and argues against Lerbekmo's (1985) magnetostratigraphic correlation. We find it difficult to envisage a scenario in which, in the absence of obvious geographical barriers, the plant that shed *Wodehouseia spinata* pollen, perhaps a member of the Rosidae, lived one half million years at the site of the present day Alberta/Saskatchewan border before migrating to the area of today's Red Deer River Valley.

The scenario is even more difficult to justify since the clastic wedge which was being shed from the rising Rocky Mountains would have caused continental environments to become established earlier in Alberta than in Saskatchewan as the Gulfian sea was withdrawing to the south.

PALEOECOLOGY

In the Maastrichtian, Alberta and Saskatchewan were part of the *Aquilapollenites* floristic province that included western Canada and Siberia, and was characterized by Lauraceae, Proteaceae, and other dicotyledonous plants (Norris *et al.*, 1975).

Several plant communities and phytosociological associations are reflected in the palynomorph assemblage reported here. Some of the major ecological groups are graphically depicted in the reconstruction of the Whitemud Flora in Figure 9. The aquatic flora is divisible into plants of lacustrine, fluvial and swamp ecosystems. Aquatic ferns such as *Azolla* and plants producing Ghoshispora megaspores could have been an integral part of all of the above habitats. Megafossil evidence also suggests that aquatic angiosperms such as lotus (*Nelumbium* sp.) and water lilies (species of *Nyphaeae*) were prevalent. The cypress swamps were perhaps dominated by species of *Taxodium* and *Nyssa*. It is noticed that species of *Aquilapollenites* are abundant in samples that contain a greater number of taxodiaceous pollen. Several members of the mistletoe family in the modern flora of the Canadian prairies are known to be parasitic on pine and spruce trees. On the other hand, in the Miocene German brown coals Teichmuller (1958) has shown plants of Loranthaceae, a family to which Aquilapollenites pollen have been compared, to be parasitic on Nyssa species. This leads us to speculate that Aquilapollenites pollen producing plants were perhaps swamp-dwellers and were growing parasitically on *Taxodium* or *Nyssa* species. A marginal vegetation of *Isoetes*, *Typha*, *Butomus* and perhaps species of *Juncus* grew on lake shores, along riverbanks and swamps. Ferns such as *Selaginella*, lycopods, *Osmunda* and cycads dominated the understory vegetation. Similarly tree ferns were an integral part of the terrestrial vegetation.

Conifers were present in the region, as can be detected from the pollen assemblage. However, it is debatable whether or not conifer forests grew adjacent to swamps, as suggested in the reconstruction of figure 9, or conifer pollen was carried by wind from distant upland forests. Several angiosperms have been recognized. The different genera of modern dicots with which the fossil pollen are compared are given in Table 2.

The Whitemud Formation contains pollen attributable to plants that can be found in modern sub-tropical cypress swamps (*e.g.* *Taxodiaceaepollenites* and *Nyssapollenites*) and pollen of plants of the temperate regions (*e.g.* *Ulmipollenites* and *Kurtzipites*). Probably, the climatic and geomorphological conditions under which the Whitemud Formation was deposited were not much dissimilar from those prevailing today in the southeastern United States.

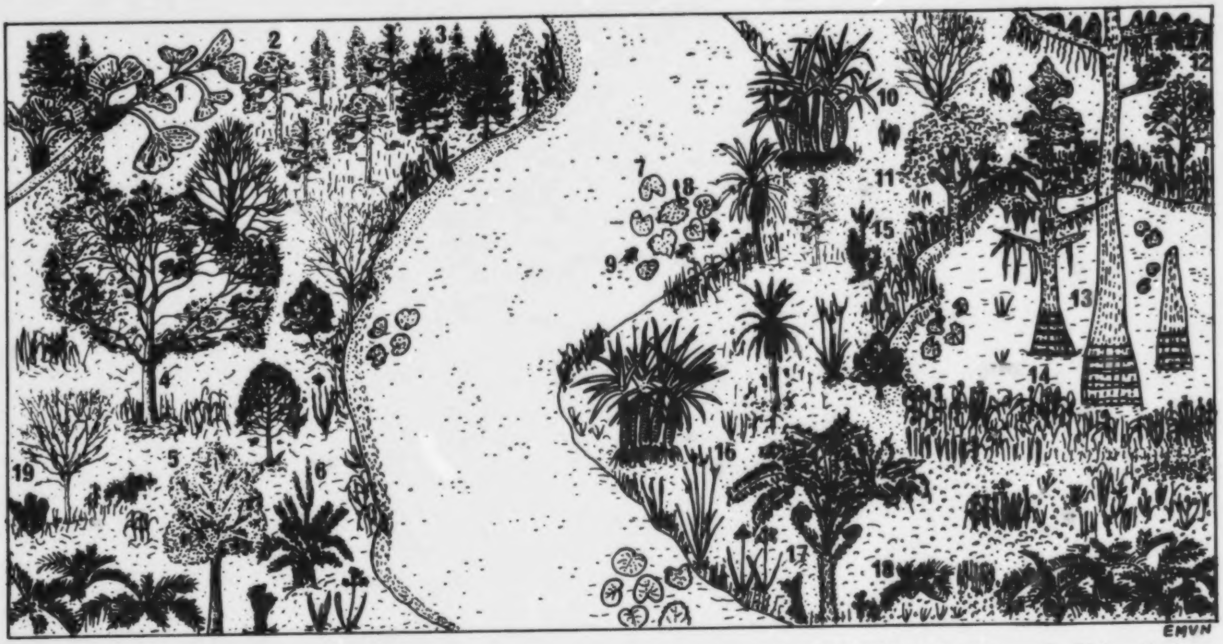

Figure 9: Reconstruction of the Maastrichtian Whitemud flora of Alberta and Saskatchewan based on plant megafossils, seed cuticles, pollen and spores. The Whitemud landscape was dominated by meandering streams, lakes and cypress swamps. Dominant plant species, numerically represented in the figure are: 1) *Ginkgo biloba*, 2) *Araucaria*, 3) unknown conifers, 4) *Phyllanthus emblica*, 5) *Cercidiphyllum*, 6) *Osmunda*, 7) *Nelumbium*, 8) *Nymphaea*, 9) *Azolla*, 10) *Pandanus*, 11) *Nyssa*, 12) *Loranthaceae*, 13) Taxodium, 14) *Typha*, 15) lycopods, 16) *Butomus*, 17) *Juncus*, 18) Cycads and 19) *Alnus* (not to scale).

ASPECTS OF DIAGENESIS

The diagenetic conditions of the Whitemud sediments are still inadequately understood. The formation is more smectitic in Alberta and more kaolinitic in Saskatchewan (Fig. 5). However, it is not clear whether or not the difference is due mainly to provenance or to diagenesis. In this paper we deal only with indications emerging from the study of siderite concretions that can be locally very abundant in the kaolinitic Whitemud.

Siderite occurs as a sub-millimetric spherulites and euhedral crystals in the Cypress Hills (Fritz et al., 1971) and as replacement of vertebrate faeces (Schmitz and Binda, 1991) or of fruits, twigs, and other vegetal remains (Nambudiri and Binda, 1989).

The Whitemud coprolites have an outer hard, compact brown or dark grey rind up to 0.5 cm thick consisting mainly of goethite (a - FeOOH) and lepidocrocite (g - FeOOH) with minor (< 10 percent) hematite and traces (< 2 percent) of quartz and siderite. A porous soft intermediate zone of variable thickness consists mainly of goethite with minor hematite and traces of quartz. The inner core, commonly lighter in colour, consists mainly of siderite and goethite with minor hematite and traces of lepidocrocite. Scanning Electron Microscope examination of the distinct zones revealed that fresh rhombohedral siderite occurs in the inner zone of the coprolites and becomes progressively more altered outward. Two different processes of mineralization can be envisaged. The faeces were either mineralized by siderite and subsequently oxidized, or alternatively, Fe-carbonate and Fe-hydroxides were precipitated simultaneously: the first in the more reducing core, the latter under oxidizing conditions on the surface. Diagenetic precipitation of siderite in fresh-water sediments has been shown by Postma (1982) to occur by mobilization of Fe^{2+} from hydroxides by decomposition of organic matter according to a reaction of the general type: $4FeOOH + CH_2O + 3H_2CO_3$ ------> $4FeCO_3 + 6HO_2$. Precipitation of siderite from interstitial waters in which Fe^{2+} and HCO_3^- were dissolved (Postma, 1981) probably occurred in the loci where faecal matter was still bacterially fermenting, thus producing CO_2 and providing porous spaces for precipitation. If, as it seems more likely, the formation of the Fe-hydroxide rind is a secondary alteration, this could have happened after all organic matter was destroyed and oxidizing water circulated in the sediment. If the precipitation of Fe-carbonate and Fe-hydroxides was simultaneous, more complex conditions analogous to recent biofilms described by Shapiro (1988) must be envisaged.

The excellent preservation of the external morphology indicates an early diagenetic mineralization of the faecal matter. Chemical and mineralogical composition suggest mobilization of iron within the sediment caused by decomposition of organic matter, with consequent precipitation of siderite as the fermentation activity of bacteria inside the faeces was still taking place. The abundance of H_2CO_3 in the diagenetic system can probably account for the non-preservation of bones and teeth.

ISOTOPIC COMPOSITION OF SIDERITE

Nine coprolites and four petrified vegetal remains from the upper part of the Whitemud Formation in the Assiniboia area of southern Saskatchewan (Fig. 1), were analyzed for $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios at the Mineralogy Institute, University of Trieste, Italy. The samples displayed various degrees of oxidation (Table 3).

The coprolites and other samples were examined by X-ray diffraction to make sure that other contaminating carbonates were absent. The samples were mainly composed of siderite and goethite with minor quantities or traces of hematite and quartz. The samples were finely ground in an agate mortar, roasted *in vacuo* at 350°C for about one hour and then reacted at 60°C with 100 percent phosphoric acid for at least 100 hours.

Table 3: Oxygen and carbon isotope ratios of siderite coprolites and vegetal remains from the Whitemud Formation in the Assiniboia area, southern Saskatchewan

SAMPLE	$\delta^{18}O$ (V-SMOW)	$\delta^{18}O$ (PDB-1)	$\delta^{13}C$ (PDB-1)	PRESERVATION
COPROLITES				
A	+19.57	-10.95	-13.13	Partly oxidized
B	-0.29	-30.22	-9.22	Oxidized
C	+24.17	-6.49	-10.77	Fresh siderite
D	+22.92	-7.70	-9.11	Partly oxidized
E	+22.85	-7.77	-12.66	Partly oxidized
F	+3.19	-26.84	-6.40	Oxidized
G	+1.64	-28.35	-4.58	Oxidized
H	+0.45	-29.50	-8.20	Oxidized
I	+5.96	-24.16	-8.62	Oxidized
FRUITS				
A	+23.28	-7.35	-7.43	Fresh siderite
C	+21.64	-8.94	-12.88	Partly Oxidized
B	+3.27	-26.76	-10.27	Oxidized
ROOT				
A	+3.95	-26.10	-7.40	Oxidized

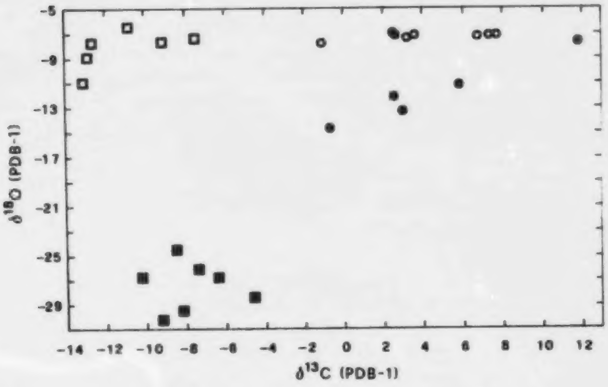

Figure 10: Plot of $\delta^{13}C$ versus $\delta^{18}O$ of siderite from the Whitemud Formation.

Key to symbols:

Squares = Coprolites and vegetal remains from southern Saskatchewan; open = fresh or partly oxidized; solid = oxidized;

Circles = Spherulites from the Alberta plains and Cypress Hills (from Fritz et al.); open = fresh; solid = oxidized.

The CO_2 evolved has been measured by means of a Finnigan Delta E mass spectrometer versus CO_2 evolved from Carrara marble (internal laboratory standard). The isotopic data were corrected for the acid fractionation effect on siderite calculated according to the relationship reported by Rosenbaum and Sheppard (1986). The results obtained are reported versus PDB-1 and V-SMOW isotopic standards in Table 3 and in Fig. 10 where the data obtained by Fritz *et al*. (1971) on siderite spherulites from the Whitemud Formation of Saskatchewan and Alberta, are also reported for comparison.

The samples measured may be divided into two groups having $\delta^{13}C$ fairly close to one another but very different $\delta^{18}O$. Those showing more positive $\delta^{18}O$ are basically formed by well crystallized siderite, while the more negative samples are essentially formed by goethite, and siderite is present only as very fine grains in trace amounts.

The large isotopic differences observed may perhaps be explained in terms of late oxidation processes which have deeply affected the structure and mineralogical composition of the coprolites and fruits.

The hypothesis which may be advanced is the following. Immediately after the deposition of the organic material, a diagenetic process took place in an environment where relatively large amounts of Fe were available along with CO_2 coming mainly from fermentation and decomposition of organic matter. The process of mineralization of the organic matter must be considered a very fast one because morphological changes of the various objects did not take place.

Taking into account a sideritization process under low temperature conditions with freshwater, the $\delta^{18}O$ values exhibited by "well preserved" siderites are compatible with the siderite-water relationship reported by Becker and Clayton (1976). Their relatively small range of $\delta^{18}O$ values may suggest that these objects retained fairly well their original isotopic composition. The comparison of this group of data with the results reported by Fritz *et al*. (1971) for "well preserved" siderite spherules show that the $\delta^{18}O$ values of the two groups of samples are very close to one another while large differences exist between the $\delta^{13}C$ results (Fig. 10). These differences may be ascribed to different environmental conditions. Fritz *et al*. (1971) explain the heavy $\delta^{13}C$ values obtained as the effect of bacterial decomposition of organic matter to form methane and carbon dioxide under nonequilibrium conditions. Subsequent isotopic exchange processes between methane and CO_2 may lead to a further enrichment in $\delta^{13}C$ of the CO_2 and, consequently, of the siderite crystals. In the case of our samples the $\delta^{13}C$ values are quite negative, even though the carbon isotopic composition of CO_2 from organic matter should be considerably more negative. One can assume that the process leading to the positive $\delta^{13}C$ observed by Fritz *et al*. took place to a lesser extent also in the case of our samples, affecting only partially the carbon isotope values. The $\delta^{13}C$ isotopic enrichment of our samples may thus be evaluated as equal to about 5 to 10 per mil (compared to mean $\delta^{13}C$ values of CO_2 from organic matter.

If we consider now the second group of samples (same $\delta^{13}C$ results but very negative $\delta^{18}O$ values) it is of importance to note that in this case only a few microcrystals of siderite are present in a goethite matrix. This seems to be the result of drastic oxidation processes which probably took place much later in presence of isotopically very light fluids, thereby causing a marked negativization of the isolated residual microcrystals of siderite. Such a process may have taken place *e.g.* throughout the Quaternary when the oxygen isotopic composition of environmental water was certainly very negative compared to Late Mesozoic water. A similar process may have taken place, at least partially, in the case of the samples studied by Fritz *et al.* which clearly show more the negative $\delta^{18}O$ values to be related to partial oxidization. It seems reasonable to assume, that a complete oxidation process would have lead in this case to a shift towards considerably more negative $\delta^{18}O(CO_3^{2-})$ values, *i.e.*, of the same order of magnitude as found in our goethite-rich samples.

Thus, isotopic ratios of siderite indicate a methanic diagenetic environment (Maynard, 1982) for the kaolinitic Whitemud Formation.

CONCLUSIONS

The Whitemud Formation of southern Alberta and Saskatchewan records the development of alluvial plains over a vast region of Western Canada during late Maastrichtian time. The fining-upwards sequence of these fluvial sediments culminated in the lacustrine deposits of the overlying Battle Formation.

Petrological and mineralogical differences in the Whitemud Formation from the plains of Alberta to the Cypress Hills and to southern Saskatchewan are probably the result of different source areas of the sediments, although diagenesis may have enhanced such differences.

The occurrence of *Wodehouseia spinata* and associated pollen taxa, together with a typical megaspore assemblage in the Whitemud Formation, strengthens the argument for a regional chronostratigraphic correlation, while arguing against a proposed magnetostratigraphic interpretation that would make the Whitemud Formation of Saskatchewan one half million years older than the Whitemud Formation of the Alberta plains.

Oxygen and carbon isotopic ratios validate the hypothesis that replacement of fossil faeces and of vegetal remains by siderite took place during early, methanic diagenesis.

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THE STRATIGRAPHY AND DEPOSITIONAL ENVIRONMENTS OF THE FOX HILLS FORMATION, BOWMAN COUNTY, NORTH DAKOTA

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ABSTRACT

The marginal marine deposits of the Fox Hills Formation (Maastrichtian) in the Williston Basin consist, along with the overlying terrestrial Hell Creek Formation and the upper portion of the underlying offshore marine Pierre Formation, of strata laid down during the penultimate retreat of the sea from the Western Interior. In the Little Missouri Badlands on the southeast flank of the Cedar Creek anticline, the Fox Hills is typically exposed as a 37 m (122 ft) thick, tabular, upward coarsening sequence consisting of (from the base): mixed or interbedded silt and clay (Trail City Member; 10 m [33 ft], offshore transition), very fine to medium hummocky-bedded sand containing a limited suite of trace fossils dominated by *Ophiomorpha* (Timber Lake Member, 19-22 m [62-72 ft], inner shelf), and a cross-bedded fine to medium-grained sand containing carbonaceous material (Colgate Member, shoreface). The Pierre - Fox Hills contact is placed at the horizon above which clay changes to silty clay, mixed or interbedded strata occur, and trace fossils appear, whereas the upper contact with the Hell Creek is placed at the base of the first substantial carbonaceous bed. Stratigraphic sequences are similar over the southwestern flank of the basin, but are half the thickness and lack the abundant marine fossil fauna and the well-developed upper sequence exposed on the southeastern flank of the basin, including the type area.

INTRODUCTION

The Cretaceous through Paleocene section in the northern Great Plains consists of large-scale cyclothems (*i.e.*, transgressive-regressive sequences) with facies aligned roughly north-south, parallel to the shoreline of the epicontinental seaway (Weimer, 1960; Hatten, 1964; Gill and Cobban, 1973; Kauffman, 1977; Reeside, 1957; Tourtelot, 1962). Offshore conditions dominated in the Williston Basin region from the Cenomanian through the Campanian with the marginal marine deposits of the Telegraph Creek (Santonian) and Judith River (Campanian) restricted to the western portion of the basin. In the Maastrichtian, marginal marine (Fox Hills Formation) and nonmarine coastal plain deposition (Hell Creek Formation) occurred over the region (Gill and Cobban, 1973) (Figs. 1, 2).

The Fox Hills Formation, recognized in the Western Interior from Colorado north to the Canadian border, is present in the subsurface and occurs in outcrop along the eastern, southern and western flanks of the Williston Basin (Fig. 1). Glacial deposits blanket the formation in the northeast. Exposures of the Fox Hills along the southwestern flank of the basin occur (Fig. 3) in the Missouri River badlands and at Fort Peck on the northern limb of the Blood Creek Syncline (Area 1); on the Poplar dome, as at south of Brockton (Area 2); at the northern end of the Cedar Creek anticline in the area south of Glendive (Area 3); and at the southern end of the Cedar Creek anticline in the Little Missouri Badlands south of Marmarth, North Dakota (Area 4). On the southeast of the Williston Basin, exposures occur (Fig. 3) in Meade County, South Dakota (Area 5); in the valleys of the Grand and Moreau Rivers in Corson, Dewey, and Ziebach Counties (Fox Hills type area) in north-central South Dakota (Area 6); and in Sioux, Emmons, and Morton Counties in south-central North Dakota (Area 7).

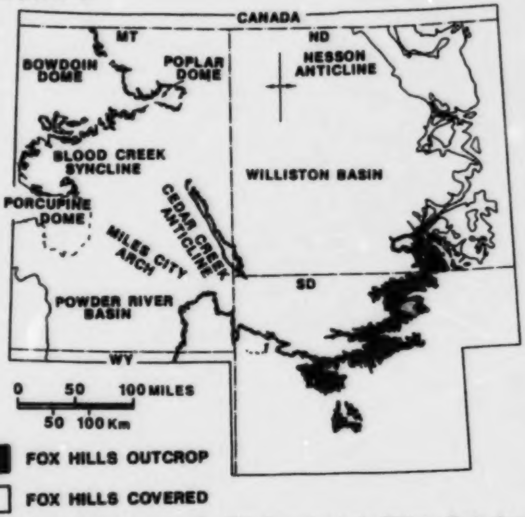

Figure 1: Areal distribution of the Fox Hills Formation in the Williston Basin (from Bluemle [1982] for North Dakota, U.S. Geological Survey [1958] for Montana, and Darton [1951] for South Dakota).

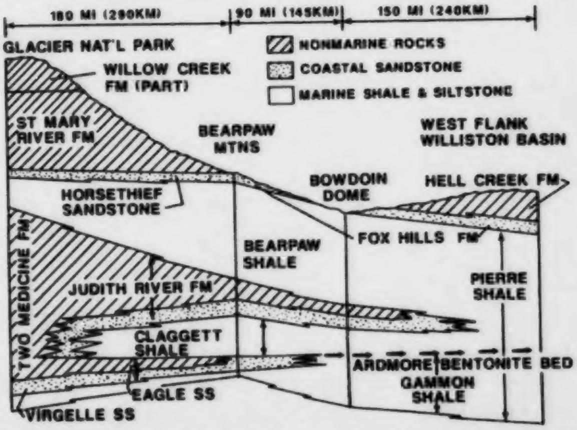

Figure 2: Stratigraphic relationships of the Fox Hills Formation on the west flank of the Williston Basin (from Rice and Shurr, 1980).

PURPOSE AND METHODS

Significant differences in thickness, fossil content, sedimentology, and depositional styles have been noted between exposures in the southeast, which include the type area of the formation and the type localities of most of its members, and outcrops on the Cedar Creek anticline, Poplar dome, and Blood Creek syncline to the west.

This paper is focused on the delineation of the Fox Hills Formation and its subdivisions and a determination of depositional environments. The area of study includes Townships 129 N-113 N and Ranges 104 W-107 W in western Bowman County, southwestern North Dakota, in the unglaciated portion of the Great Plains Physiographic Province (Fig. 4). The towns of Marmarth and Rhame are situated to the north and northeast on U.S. Highway 12. The region is dissected by the north-flowing Little Missouri River, bridged by U.S. 12 at Marmarth.

PREVIOUS WORK

The Fox Hills Formation was defined by Meek and Hayden (1861) from exposures in north-central South Dakota on the southeastern flank of the basin. Subsequent investigations can be loosely grouped into three periods: 1) 1900-1940, recognition and description of the formation in the region as part of the United States Geological Survey coal field evaluation program; 2) 1940-1960, subdivision of the formation by the Geological Surveys of North and South Dakota; and 3) 1960-present, a reassessment of the formation definition and its subdivisions, including focused studies on stratigraphy, paleontology, and sedimentary structures. Distribution of select study areas is shown on Figure 3.

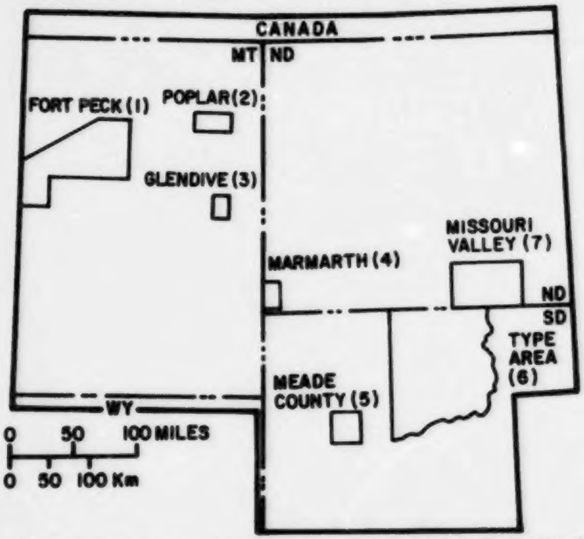

Figure 3: The distribution of select outcrop areas for the Fox Hills Formation in the Williston Basin. By area, previous work includes Areas 1 to 6.

Accounts of the conceptual evolution of the formation and its subdivisions are provided in Waage (1968) for the type area and Feldmann (1967, 1972) for the adjacent area of North Dakota. Brown (1962) and Frye (1967, 1969) documented the development and resolution of the "Cretaceous Boundary" or "Laramie" problem, which involved the upper contact of the Fox Hills in the early part of the century. The historical development of the formation and its subdivisions on the southwest flank is provided by Daly (1984).

In 1968, Waage expanded the type area and, within that region, redefined and described the formation and its subdivisions and determined depositional environments. Reports on the southeastern rim have focused, in large part, on the rich fossil fauna and paleoecology of the Fox Hills. Key studies include Speden (1970) for bivalves in the type area and Feldmann (1967, 1972) for bivalves, and Erickson (1971, 1974) for gastropods. Additional studies include Palubniak (1972) for oysters and Artzner (1974) for pollen.

Pettyjohn (1967) described the atypical coal-bearing Fox Hills strata of the Meade County area in South Dakota.

With the lack of marine body fossils on the southwestern flank, surface investigations there have focused on determining stratigraphic relationships and depositional environments based on sedimentology, petrography, and sedimentary structures (Butler, 1980; Lepp, 1981; Crowder, 1983; Wheeler, 1983; Wilde, 1984; Daly, 1984).

In the North Dakota portion of the basin, knowledge of the formation is based on a structure map on the Pierre Formation (Carlson, 1982b), correlations by the North Dakota Geological Survey (Carlson, 1979, 1982a, 1983a, 1983b; Bluemle, 1984), and Rehbein (1977) and Rice (1976, 1977) and a statewide review of the formation (Cvancara, 1976).

Crowder (1983) studied the outcrops in Areas 1, 3, the southwestern extent of the Cedar Creek anticline, the northern and western flanks of the Black Hills, and in the Casper, Wyoming, area. Studies outside the Williston Basin include Weimer and Tillman (1980) in the Denver basin and Land (1972) on the Rock Springs uplift in southwestern Wyoming.

The descriptions in the early 1900s came out of United States Geological Survey coal field studies focused on overlying coal-bearing strata (Leonard, 1906, 1908; Calvert, 1912; Winchester *et al.*, 1916; Hares, 1928), or on the "Cretaceous Boundary" problem (Stanton, 1910; Thom and Dobbin, 1924; Dobbin and Reeside, 1929). Recent reports include Feldmann (1967, 1972), and Carlson (1979).

GEOLOGIC SETTING

The Williston Basin, a northwest-southeast trending stratigraphic and structural intracratonic basin (Fig. 1), contains a maximum 4,612 m (15,128 ft) of sediments of Cambrian to Recent (Carlson and Anderson, 1970). Dips average 3.8 m/km (20 ft/mile) toward northwestern North Dakota. The broad Miles City arch separates the Powder River Basin from the Williston Basin. Two major structural highs occur in the basin, both in North Dakota (Gerhard *et al.*, 1982): the north-trending Nesson anticline to the north, and the northwesterly oriented Cedar Creek anticline, in the southwest. These and other structures, including the Poplar dome in the northwest, represent the surface expression of episodic basement block movement (Anna, 1986).

The Cedar Creek anticline controls the outcrop pattern in the study area, bringing the uppermost Cretaceous to the surface in a 9 to 32 km (6-20 mile) wide outcrop band (Figs. 1, 4). The anticline 1) strikes N 30 E and plunges to the northwest, 2) has steeper dips on the southwest flank (4-40°) than on the eastern

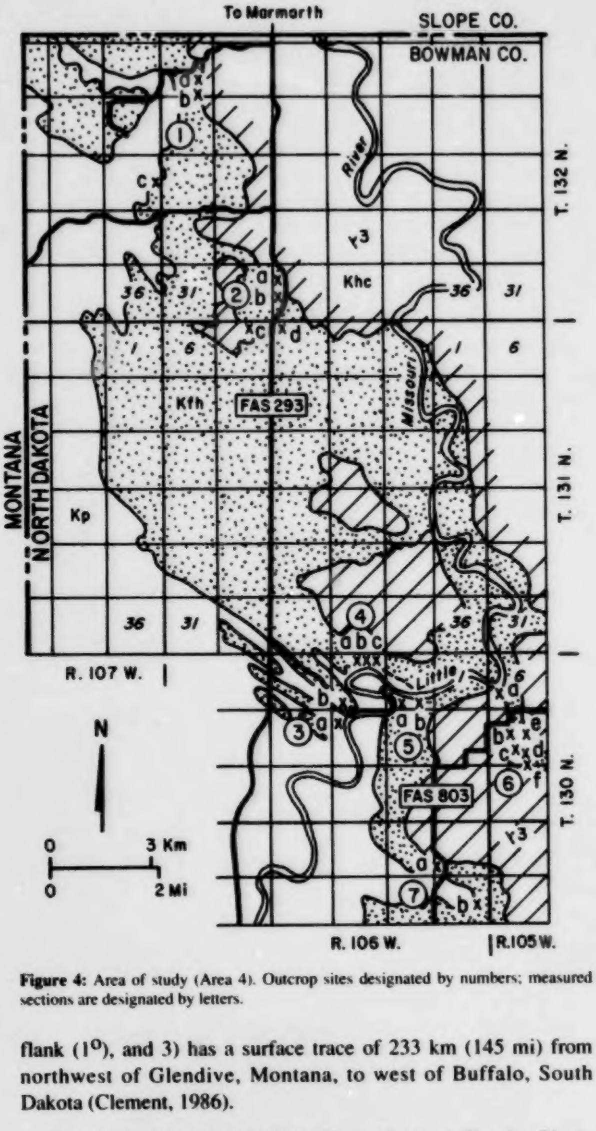

Figure 4: Area of study (Area 4). Outcrop sites designated by numbers; measured sections are designated by letters.

flank (1°), and 3) has a surface trace of 233 km (145 mi) from northwest of Glendive, Montana, to west of Buffalo, South Dakota (Clement, 1986).

The Maastrichtian Fox Hills Formation overlies the Pierre Formation (Bearpaw Shale of central Montana) and is overlain by the Hell Creek Formation. The Fox Hills and Pierre form the Montana Group (Eldridge, 1889). Fox Hills correlatives for the Western Interior include the Lennep and Horsethief Formations in Montana (Gill and Cobban, 1973, p. 20), the Dad Sandstone Member of the Lewis Shale in Wyoming (Gill and Cobban, 1973, p. 20), the Lion Canyon Sandstone of the Williams Fork Formation and the Trinidad Sandstone in Colorado (McGookey *et al.*, 1972), and the Boissevain Formation in southern Manitoba (Williams and Burke, 1966, p. 169, 174-5).

FORMATION DEFINITION AND SUBDIVISION

Meek and Hayden (1861) defined the Fox Hills Formation (Formation 5 in their earlier publications) as an unspecified thickness of "grey, ferruginous, and yellowish sandstone and arenaceous clays" (p. 427), that cropped out along the Cheyenne-Moreau divide in north-central South Dakota. The formation was considered the uppermost marine (and in the thinking of the time, therefore, uppermost Cretaceous), by its extensive marine fauna, and included the strata between the homogeneous marine Pierre Shale (formerly Formation 4) and the lignite-bearing, nonmarine, Great Lignite Group (Tertiary). The conformable Pierre - Fox Hills contact was placed where the clay changed upsection to sandy clays. The upper contact was chosen at the base of Bed Q, a sand believed by Meek and Hayden to mark the onset of brackish conditions.

Waage (1968 p. 60-61), following the original concept of Meek and Hayden in the type area, placed the basal contact of the Fox Hills at the obvious increase of silt in the section. In several areas, "sandstone" has been substituted for "formation." This practice (*e.g.*, Stanton, 1910) placed the lower contact at the base of the first sand, thereby relegating the transition beds to the Pierre. Waage (op. cit.), objected to this "Rocky Mountain" definition. Currently the contact is typically placed in the transition zone (Fig. 6).

Stanton (1910) demonstrated the marine origin of Bed Q, and Calvert *et al.* (1914) included Bed Q in the Fox Hills. This criterion, reiterated in North Dakota by Laird and Mitchell (1942), has been used by subsequent workers. Thus, the uppermost Cretaceous section on the southwestern flank of the Williston Basin comprises upward 1) marine shale; 2) interbedded marine shale, silt, silty shale, and sandy silt (transition zone); 3) marine silty sand to sand; and 4) nonmarine sands, silts, clays, lignites, and carbonaceous zones. The Fox Hills section, from the base of the interbedded deposits to the base of the nonmarine strata, is typically 35 to 46 m (115-150 ft) thick. To the southeast, including the Fox Hills type area, additional interbedded sand and clay, cross-bedded sands, and a regionally continuous indurated sandstone occur in the upper part of the formation, resulting in a thickness for the Fox Hills of over 91 m (300 ft).

Eight members are currently recognized in the Fox Hills of the Williston Basin (Fig. 7) and, with the exception of the Colgate, all were defined on the southeastern flank of the basin. The name Colgate was given to the distinctive white sandstone on the Cedar Creek anticline (Dobbin and Reeside, 1924), and was the only member name in wide use until the 1940s. A complete subdivision of the formation into members (O-Y) Trail City, Timber Lake, Bullhead, and Colgate, emerged from the work of the South Dakota Geological Survey (Morgan and Petsch, 1945; Curtiss, 1955; Stevenson, 1956). Waage (1968) subdivided the strata in the type area and vicinity as follows (Figs. 6, 7): 1) a wedge of silty to sandy clay (Trail City Member), that thins from 60 m in the southwestern corner of the type area to 10 m or less in the Missouri Valley of North Dakota; 2) a north-northeastward thickening wedge of sand (Timber Lake Member), that extends from the central portion of the type area and increases in thickness at the expense of the underlying strata, to over 50 m in the Missouri Valley of North Dakota; 3) an upper unit (Iron Lightning Member), consisting of interbedded clay, sand, and silt (Bullhead lithofacies) and a stratigraphically variable sand (Colgate lithofacies), that thins northward from 50 m in the southwestern corner of the type area to 30 m in the Missouri Valley of North Dakota. Klett and Erickson (1976) designated a 2 to 5 m thick sand capping the section in Emmons, eastern Sioux, and Corson Counties as the Linton Member.

The informal term "lithofacies," introduced by Waage (1968), is defined as follows (p. 46): one or more bodies of sediment or sedimentary rock, distinguished from enclosing deposits by noteworthy lithologic, organic and/or structural characteristics. This usage serves to single out lithologic units essential for environmental reconstruction. Further, the formation was informally divided (*e.g.*, Waage, 1968; Cvancara, 1976) into lower (Trail City and Timber Lake) and upper (Iron Lightning and Linton) on the southeast flank.

In the southeast, the Trail City Member is a sandy shale containing three to five locally continuous zones of fossiliferous concretions (Little Eagle lithofacies) which in the type area is divided into two lithofacies: the Irish Creek in the southwest and the Little Eagle in the northeast and east. The Irish Creek lithofacies consists of an upper half, laterally equivalent to the Timber Lake Member, which is composed of massive clay, silt, and very fine sand, and a lower half, equivalent to the Little Eagle lithofacies, of massive light grey clay, silt and very fine sand, and interbedded clay and silt, and three to five locally continuous zones of fossiliferous concretions. The name Trail City has been used in Areas 6 and 7 (Erickson, 1971; Bluemle, 1984), in Area 4 and tentatively in Area 5.

In the southeastern area, the Timber Lake Member is composed of poor to well sorted, fine to medium sand that weathers to a yellow or orange brown. Fossiliferous concretions occur near the base. The upper half of the member, the *Tancredia-Ophiomorpha* biofacies, is characterized by abundant ferruginous beds of mudstone or sandstone, trough and tabular cross stratification, and abundant ferruginous *Ophiomorpha* burrows. The basal sands are typically glauconitic. Regionally, the member is recognized in Areas 3, 4, 6 (eastern half), and 7.

The Iron Lightning Member, only exposed in the southeast, contains the Bullhead and Colgate lithofacies. The Bullhead is composed of 3 to 30 cm thick interbeds of fine to medium, white to grey sand and silt, and brown to grey clay. The bedding is often contorted. Biogenic remains include fecal pellets and plant fragments that are concentrated along bedding planes, as well as *Lunatia* and *Dentatium* indicating marine affinities. In the type area, the Colgate is composed of light-grey weathering, very fine to medium, lithic sand. Tabular and trough cross-bedding are common, with the lower bounding surfaces accentuated by clay or carbonaceous layers. *Corbicula* and *Crassostrea* indicate brackish to freshwater affinities. Plant debris is common, along with 0.3 to 4 m diameter, ovoid concretions that often form ledges. The Iron Lightning is recognized in Areas 6 and 7. The Bullhead is absent to the west of Area 6, while on the southwest flank, the Colgate is treated as a member.

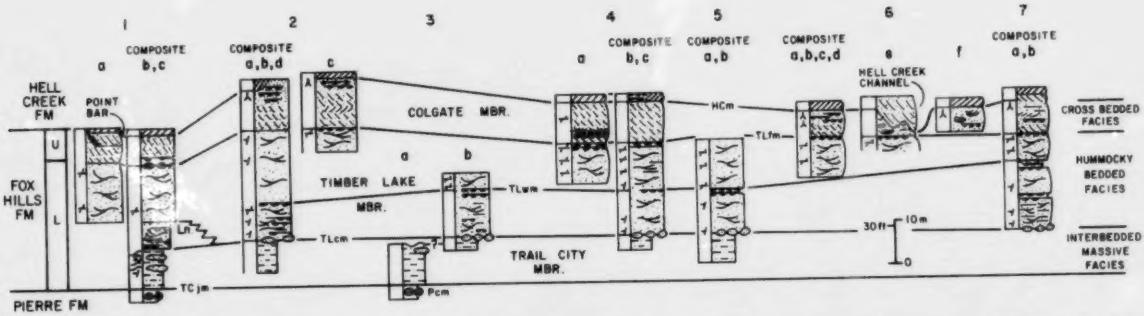

Figure 5: Stratigraphic correlation for measured sections in the study area. Refer to Figure 6 for Legend; refer to Figure 4 for site and measured section locations. "Ln" designates lenticular subfacies. Marker zones are designated as follows (explanation in text): Pcm=Pierre cone-in-cone marker; TCjm=Trail City jarosite marker; TLcm=Timber Lake concretion marker; TLwm=Timber Lake wavy marker; TLfm=Timber Lake ferruginous marker; and HCm=Hell Creek marker.

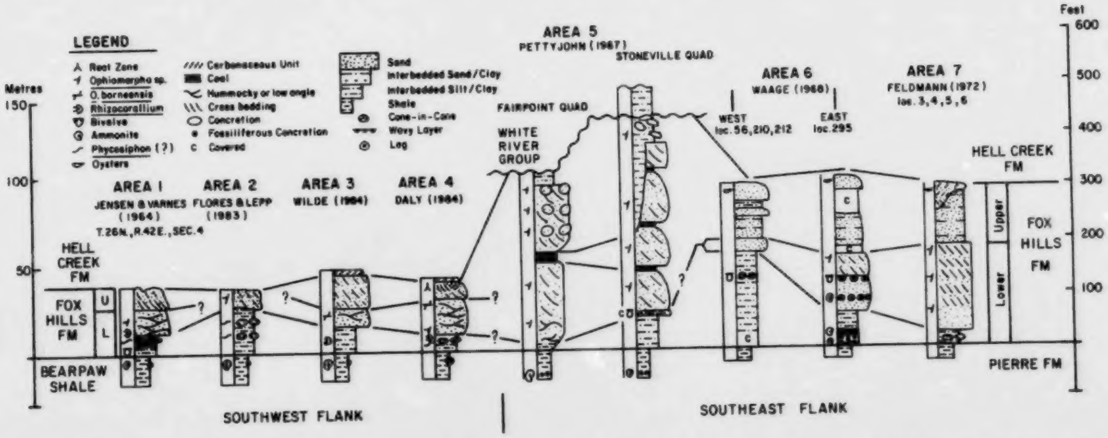

Figure 6: Representative outcrop sections of the Fox Hills Formation around the southern flank of the Williston Basin. Outcrop areas are shown on Figure 3. Sections are composites unless otherwise noted.

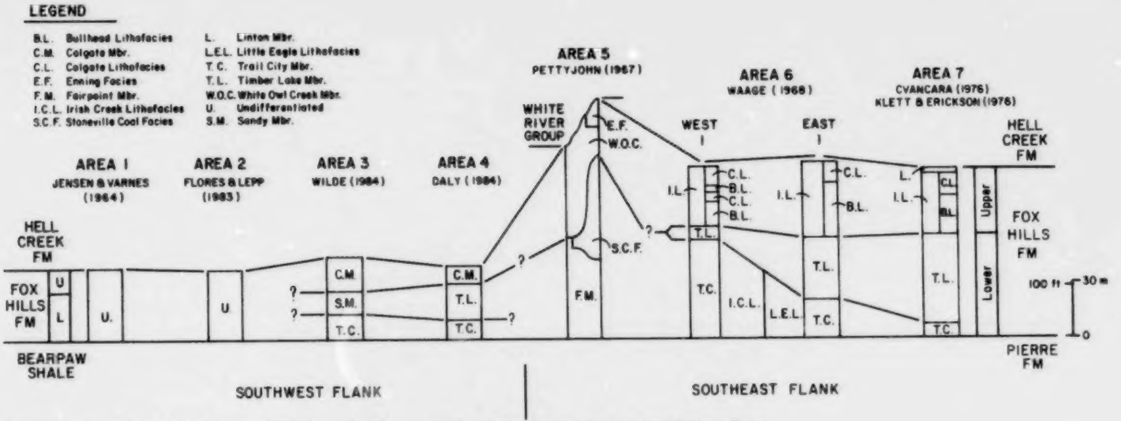

Figure 7: Stratigraphic subdivision of the Fox Hills Formation in outcrop around the southern flank of the Williston Basin.

The Linton Member is composed of light olive-grey to grey-brown, fine grained, moderately to poorly sorted, indurated siliceous sandstone. The unit is generally massive, but contains some horizontal and trough cross-stratification. Roots, wood fragments, *Equisetum* and *Ophiomorpha* occur in this butte-capping, sheet sandstone that lies above the Iron Lightning in Area 7.

In Area 5, Pettyjohn (1967) divided the Fox Hills into the Fairpoint and White Owl Creek members. The Fairpoint Member consists of a lower shale, silt, and sand sequence, and middle sandstone sequence, and an upper coal-bearing sequence (Stoneville Coal Facies). *Ophiomorpha* occurs throughout. The Owl Creek Member is composed of silt, sand, sandstone, and, at the top, purple clay (Enning Facies). *Ophiomorpha* occurs in the sands of the lower portion of the unit. These names are applicable only in Area 5.

In the west, Feldmann (1967, 1972) used the names Timber Lake and Colgate in Area 4, and Carlson (1979) the Trail City. In Area 3, Butler (1980) divided the section into Timber Lake and Colgate (the usage followed by Vuke-Foster *et al.*, [1986]), whereas Wilde (1984) divided the section into Trail City, "sandy member," and Colgate. The strata in Areas 1 and 2 have not been subdivided.

In the study area, Leonard (1908) tentatively designated 24 to 26 m (79-85 ft) of unfossiliferous strata exposed along Little Beaver Creek (western Bowman County, Little Beaver Creek Sites 1a, 1b; Figs. 4, 5, 8) as the Fox Hills Formation. He found no fossils to confirm the strata as marine, but based the designation on similarity in stratigraphic position and appearance to outcrops of the Fox Hills in the vicinity of Hell Creek, southwest of Fort Peck, Montana. From the base, the section (summarized from Leonard, 1908, p. 44) is as follows: clay, sandy, finely laminated 6 to 8 m (20-26 ft); ledge-forming sandstone, yellow 3 m (10 ft); and sandstone, light greenish grey, massive 15 m (49 ft). The Pierre-Fox contact is below the surface of the creek at this location.

Stanton (1910) demonstrated the marine origin of the strata in descriptions of the Pierre through Hell Creek section on Little Beaver Creek southwest of Marmarth. He examined the upper contact and the lower Hell Creek section, noted that although the change from marine to fresh water conditions seemed abrupt, it was actually transitional as shown by the persistence of brackish water sediments in the basal Hell Creek, and concluded that unconformities were local and of minor significance.

The Pierre-Fox Hills contact in the study area (Sites 1, 2d, and 5a) was first described by Hares (1928), who drew the contact at the base of the lowest massive sand, thereby including any interbedded silt and clay in the Pierre. Feldmann (1967, 1972) also placed the lower contact at the lowest sand (Site 2d; Figs. 8 and 9), but noted a 3 m (10 ft) transition zone that he included in the Pierre. Carlson (1979) included the transition zone in the Fox Hills. Elsewhere on the southwestern flank, the contact has typically been placed at the change upsection from the homogeneous shale of the Pierre to interbedded clays, silts, and muddy sands (*e.g.*, Jensen and Varnes [1964] in Area 1, Colton [1967], and Lepp [1981] in Area 2, and Thom and Dobbin [1924] and Butler [1980] in Area 3). In contrast, Crowder (1983) and Wheeler (1983) placed the contact at the base of the first sand. The upper contact of the Fox Hills corresponds to the top of the marine section and was placed by Hares (1928) and Dobbin and Reeside (1929) at the colour change from the bright white sand of the Colgate to the drab of the Hell Creek. Frye (1967, 1969) and Feldmann (1967, 1972), after Laird and Mitchell (1942) picked the upper contact at the base of the first substantial carbonaceous zone.

The distinctive bright white sand of the Colgate and was the only member recognized in the southwest rim until Feldmann (1967, 1972) defined the Timber Lake in Area 4. Thom and Dobbin (1924) designated the Colgate, a unit they believed was the highest marine strata in the Cretaceous, a member of the Fox Hills in Area 3, restricting it to the 12 m-thick "conspicuous white, upper sandstone of the Fox Hills" that contained both *Halymenites (=Ophiomorpha) major* and leaf imprints in exposures around the anticline. Leonard (1906) and Calvert (1912) had considered these beds, along with the remainder of the current Fox Hills,to be part of the Lance. Using the "white" colour as the diagnostic criterion, Hares (1928) introduced the Colgate to the study area (Sites 1a, 1b, 2d, and 5a; Figs. 4, 5, 8), noting that both the Colgate and the unnamed strata below contained *Halymenites (=Ophiomorpha) major*. Dobbin and Reeside (1929) assigned the entire Fox Hills section (Site 1b) to the Colgate.

Feldmann (1967, 1972) described the Fox Hills section as a 24 m thick sand sequence (very fine to medium sand) and divided the section into the basal 14 m thick Timber Lake Member, and the overlying Colgate Member, and noted a 3 m thick transition with the underlying Pierre. At Site 2d, he described the Timber Lake as 14 m of buff to yellow, fine to medium sand, which contained chocolate-brown shale interbeds at the base and cross-bedding and discontinuous, ferruginous sandstone ledges in the upper half. The Colgate was described as 12 m of "dirty, white sandstone" containing lignitic shale partings and exhibiting fluted weathering. The weathering pattern and lignitic partings were considered diagnostic. Carlson (1979) concurred, but assigned the transition zone to the Trail City.

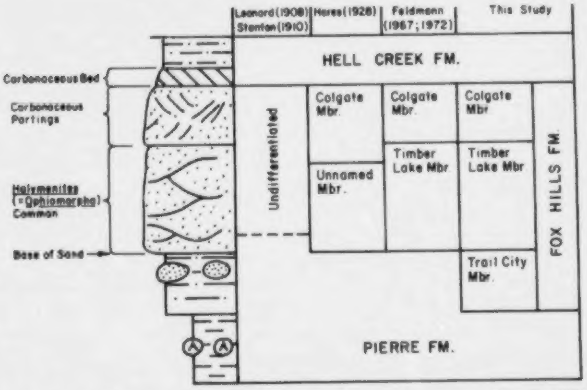

Figure 8: Historical development of the definition and subdivision of the Fox Hills Formation in the study area (Area 4).

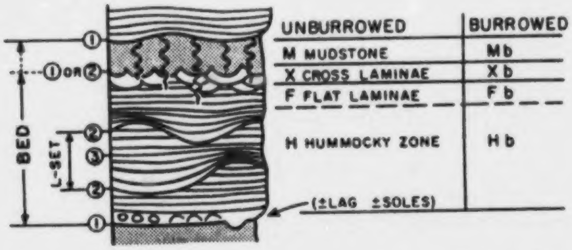

Figure 9: Idealized hummocky sequence with notation for component zones (from Dott and Bourgeois, 1982). In the study area, W zones (wavy bedding) occur in place of M zones. Note the distinction between hummocky lamina sets and hummocky beds. Numbers refer to 1=first order bed to bed contacts, 2=second order bedform to bedform contacts, 3=third order lamina to lamina contacts.

SEDIMENTOLOGY

The Fox Hills Formation in the study area (Area 4; southwestern rim of the Williston Basin) is a tabular unit, typically 37 m (121 ft) thick, that consists of a basal 10 m (33 ft) silt-clay unit, and an overlying sand unit 27 m (88 ft) thick (Fig. 5). The Pierre-Fox Hills contact is placed at the change from shale to silt, thereby including the transition zone in the Fox Hills.

On the basis of inorganic sedimentary structures, three facies can be delineated (from the base): interbedded-mixed, hummocky-bedded, and cross-bedded. The hummocky-bedded facies contains three subfacies (from the base): lenticular, tabular, and amalgamated.

The three members correspond to the sedimentary structure facies as follows (from the base) (Fig. 5, 8): Trail City (interbedded-massive), Timber Lake (hummocky-bedded), and Colgate (cross-bedded). Furthermore, the section is divided into lower (Trail City and Timber Lake) and upper (Colgate) Fox Hills.

SEDIMENTARY STRUCTURES

Major inorganic sedimentary structures include wavy bedding, hummocky bedding, and cross bedding. These bedforms are typically restricted to the sandy portion of the formation.

Wavy bedding is a composite heterolithic bedset characterized by Reineck and Singh (1975, p. 98-99, 103) as: 1) alternating, mud drape and sand ripple layers that are horizontally continuous and vertically isolated; 2) mud drapes that almost completely fill the ripple trough and make a thin cover over the ripple crest, thereby forming a nearly horizontal upper surface; and 3) many alternating sand ripple and shale drape couplets. Lenticular bedding resembles wavy bedding except that the sand ripple layers are discontinuous (Reineck and Singh, 1975, p. 100-1). Wavy bedding is a major component of the lenticular subfacies. It forms W zones at the top of hummocky beds in the tabular-bedded and amalgamated subfacies, and occurs in the cross-bedded facies. The wavy beds typically feature wave (oscillation) ripples. Wavy layers, interlayered wave ripples and drapes represent a minor fluctuation in a low-energy regime. Wave ripples form in low-energy oscillatory conditions, while drapes are suspension deposits formed during slack water periods.

Hummocky L-sets occur in a relatively consistent, heterolithic, composite bed sequence termed "hummocky bedding" (Dott and Bourgeois, 1982) (Fig. 9). Hummocky beds are bounded at the base by sharp erosional planar contacts marking the bases of hummocky zones (H zones). The H zones are composed of individual hummocky forms termed L-sets. These zones are overlain by planar laminae (F zone) over which lies a zone of ripples (X zone). In the study area, the sandy HFX sequence is overlain by wavy layers (wave ripples and drapes) or wavy bedding, here termed W zones. Elsewhere, in Area 1 for example, hummocky beds are overlain by mudstones (M zones). The F and X zones are continuous extensions of the H zone; no contact separates these layers. W zones may be continuous with the underlying bed or separated from it by a scour surface. Stacked hummocky beds form the strata of the hummocky-bedded facies, and tabular isolated sets occur in the underlying facies. Hummocky bedding (HFX) represents a single, short duration sedimentation event, but the wavy bedding occurs between depositional events.

The sandy portion of the formation (hummocky-bedded and cross-bedded facies) consists of mainly muddy, sublithic or subarkosic, very fine to medium sand. Very fine sandy mud occurs at the base of the sand section. The lower portion of the tabular-bedded subfacies is glauconitic. The following generalizations can be made with respect to the texture of the sand units: 1) the percent of matrix decreases upsection through the hummocky facies from 50 to 10 percent, whereas the cross-bedded facies contains 15 to 30 percent matrix; 2) the sediments of the hummocky-bedded facies are moderately sorted, whereas those of the cross-bedded facies are moderately to poorly sorted; 3) the median grain size increases upsection from 4.0 phi (base of the hummocky facies) to 2.80 phi (top of the hummocky-bedded facies) to 2.30 phi (cross-bedded facies); 4) lateral variability in texture is low in the hummocky-bedded facies, but relatively high in the cross-bedded facies; and 5) the grain-size distributions exhibit a strong positive skewness throughout the section.

Although Stanton (1910) reported *Leda (Yoldia) evansi, Tellina scitula, Entalis(?) paupercula,* and *Halymenites (=Ophiomorpha) major* from the vicinity of Site 1, only leaf remains, shark teeth, and ichnofossils were observed by the writer. There are dwelling traces, including a single example of *Thalassinoides* (?) and numerous *Ophiomorpha* (*O. nodosa* and *O. borneensis*); the grazing *Phycosiphon*(?); and unidentified feeding structures. Leaves, notably *Dryophyllum*(?), horsetail stems (*Equisetum*), and root molds are found at or near the top of the formation.

Ophiomorpha, which represents the burrows of shrimp, is the most plentiful and widely distributed trace fossil in the formation. *O. nodosa* is characterized by burrows with an external diameter of 5 to 7 mm (0.21-0.3 in) and a pelletal lining 1 mm (0.04 in) thick, that is composed of 1 mm (0.04 in) diameter pellets or have a pelletal lining 3 mm (0.12 in) thick, and pellets 2-4 mm (0.1-0.2 in) in diameter. The burrows form in irregular boxworks (Frey *et al.*, 1978) 10 to 20 cm (4-8 in) thick.

O. borneensis is characterized by exterior burrow diameters of 2.5 to 4 cm (1-2 in). In cross section the tunnels consist of an orange-brown, 4 to 7 mm (0.2-0.3 in) thick, indurated lining of bilobate pellets and an interior filling of loose or indurated sand, either massive, or arranged in meniscate laminae. The pellets are lozenge-shaped, vary from 3 x 5 mm (0.1 x 0.2 in) to 5 x 10 mm (0.2 x 0.4 in) and are usually arranged with their long axis normal to tunnel direction. The interior lining is smooth.

Tunnels and galleries are arranged in tiered mazes 0.3-0.6 m (1-2 ft) thick (Frey *et al.*, 1978, Fig. 2F); horizontal components (galleries) are abundant but vertical elements (shafts) are rare. Vertical elements connect maze tiers and remnant shafts, 2 to 3 cm (0.8-1.2 in) in height, occur atop maze systems. Fecal pellets lie on hummocky laminae or in laminated burrow interiors. In the horizontal plane, tunnels radiating from shafts give the appearance of rimless wagon wheels. Each "wheel" can cover 1 to 4 m^2 (3-13 ft^2) in area.

STRATIGRAPHY

MARKER ZONES

Six marker zones, labeled below with the capital letter of the host member or formation, followed by a small letter "m," are defined for the Fox Hills section in the study area (Fig. 5).

The Pierre cone-in-cone layer (Pcm) marks the highest occurrence of cone-in-cone concretions in the Pierre Formation. The concretions average 20 cm (0.6 ft) in diameter and exhibit cone-in-cone structure throughout. The marker was observed at Sites 1c and 3a. The Pierre-Fox Hills contact typically occurs 1-3 m (3-10 ft) above this zone.

The Trail City jarosite marker (TCjm) is composed of moderate yellow (5Y 7/6), jarosite-impregnated silt interbedded with grey clay. The zone is 1.5 m (5 ft) thick at Site 1c, but occurs as stringers at Site 3c. The base of the TCjm coincides with the Pierre-Fox Hills contact.

The Timber Lake concretion marker (TLcm) is a zone of spherical to elliptical, light grey, very fine sandy silt concretions averaging 0.3-0.5 m (1-2 ft) in diameter. The base of the zone marks the Trail City-Timber Lake contact at all outcrops. The zone is a partly mixed hummocky bed, as evidenced by remnant hummocky lamination.

The Timber Lake wavy marker (TLwm) is the highest occurrence of a substantial wavy-bedded layer (W zone) associated with tabular-bedded hummocky strata. The thickness of this marker ranges from 0.15 m (0.5 ft) (Site 1b) to 0.3 m (1 ft) (Site 2a) to 1.2 m (4 ft) in the southern outcrops. The zone proves helpful at outcrops where only the upper portion of the hummocky-bedded facies is exposed. This marker is planar and can be laterally continuous for 0.5 km (0.3 mi).

The Timber Lake ferruginous marker (TLfm) consists of a planar, ferruginous, wavy layer or bed. The top of this marker corresponds to the top of the Timber Lake Member, the hummocky-bedded facies, and, typically, the uppermost occurrence of *O. borneensis*-zoned bioturbation.

The Hell Creek marker (HCm) occurs at the base of the Little Beaver Creek Member of the Hell Creek Formation and consists of indurated dark yellow brown (10YR 4/2), carbonaceous, muddy, fine to medium sand that may be massive (Sites 1, 2, 7a) or exhibit trough-bedding (Sites 6b, 6c, 6d, 7a). The HCm is typically 1 m thick, but may be thin or absent (Site 2b, 4, 6c-d). The base of this unit marks the Fox Hills-Hell Creek contact as indicted by Feldmann (1967, 1972), Frye (1967, 1969), and Carlson (1979).

UNITS

Pierre-Fox Hills Contact

The Pierre Formation is a typically dark grey weathering, massive clay or silty clay, with blocky fractures, and a surface exposure covered with indurated shale "chips" and translucent selenite. The Pierre forms rounded, gentle slopes, is slick when wet, and is characterized by dendritic drainage. Within a few metres of the contact, the Pierre becomes more silty and bedded thin silt stringers.

The Pierre-Fox Hills contact is placed where the homogeneous, grey clay of the Pierre is replaced upsection by the mottled, mixed, light grey (N7) silt and grey clay or by the thinly intercalated silt and clay of the interbedded-massive facies. Where the contact is exposed (Sites 1c, 3a), a break in slope and jarosite stringers are present. A layer of cone-in-cone concretions (Pcm) occurs 1-3 m (3-10 ft) below the contact.

Fox Hills-Hell Creek Contact

The strata of the Little Beaver Creek Member of the Hell Creek Formation (Frye, 1969, p. 20-21, 61-63) consist of 10-12 m (33-39 ft) of tabular, interbedded claystone and mudstone layers, 0.15-0.5 m (6-20 in) thick, that contain root molds and leaf hash and minor sand beds. This member is overlain by the medium sand and large scale trough cross-beds of the Marmarth Member. The typical colours are browns and greys, and bentonitic beds are common.

The Fox Hills-Hell Creek contact is placed between the basal carbonaceous sand of the Little Beaver Creek Member (HCm) and the underlying cross-bedded facies of the Fox Hills Formation. The contact as here defined is the same as that delineated by Feldmann (1967, 1972), Frye (1967, 1969), and Carlson (1979) (Figs. 5, 8).

The contact typically occurs atop a scoured surface and may be planar or channeled. At most outcrops (Sites 1, 2, 6, 7), the upper 2 m (7 ft) of the Colgate consists of light olive grey, fine-medium, muddy sand (70-85 percent sand) that contains root molds. The upper surface of the member is scoured into broad, shallow troughs (1-2 m [3-7 ft] relief) that are overlain by the HCm.

However, at Site 6e, this sand bed is locally removed by Hell Creek channeling, and the contact is an irregular surface with significant relief (3-6 m [10-20 ft]). At Site 4, the HCm is missing over part of the area, and HCm cobbles occur as lag in the underlying Fox Hills. Overall, the scouring and channeling are of local extent and do not appear to represent a major uncomformity.

Trail City Member

This interbedded-massive facies contains two components: tabular, sandy-silt hummocky layers, and interbedded or mixed silt and clay. The facies is up to 10 m (33 ft) thick with an 8 m (26.4 ft) section exposed at Site 1c. The massive component consists of a mottled, mixed, light grey (N 7) silt and grey clay. Remnant, finely intercalated silt and clay layers occur in some exposures in the upper portion of the facies; bedding planes are gradational. The mixed and mottled character of the sediment strongly suggests biogenic reworking, although no identifiable traces were found. Mixed sandy-silt concretions may represent reworked hummocky beds.

In the hummocky component two or three tabular hummocky beds (HbMb), 0.5 m (1.6 ft) thick and composed of light grey (N 7), very fine, sandy silt, occur in the upper portion of the facies at Little Beaver Creek (1c). The beds have sharp basal contacts, are laterally continuous in outcrop, and are bioturbated in their upper portions by *O. nodosa* and *Phycosiphon*(?). The hummocky L-sets of the H zone are small-scale.

The texture and sedimentary structures of the Trail City in the study area are similar to those in the type area and most closely resemble the strata of the Little Eagle lithofacies. At the type locality of that lithofacies, tabular, 0.3-0.5 m (1-2 ft) thick hummocky beds occur interbedded with mixed silt-clay and fossiliferous concretions. The absence of the fossiliferous concretions in the study area does not rule out designating the transitional beds as the Trail City Member. In the type area the concretions are laterally continuous between the Trail City and Timber Lake.

Timber Lake Member

The tabular hummocky-bedded facies is typically 19-22 m (63-72 ft) thick. The basal contact, a scour surface at the base of a hummocky bed, is typically drawn where hummocky beds become tabular and stacked without intervening mixed zones. The contacts between the subfacies are conformable, and the upper contact of the facies is erosional. Internally, grain size and sand content increase up section; the sediment is moderately sorted, and lateral variation in sorting is low.

Three subfacies — lenticular, tabular and amalgamated — can be delineated on the basis of the ideal internal sequence of the hummocky beds.

The lenticular subfacies is 2-3 m (7-10 ft) thick and was observed only at the Little Beaver Creek exposures (Sites 1b, 1c). The subfacies is characterized by L-set-lenses (HFX), 0.5 m (1.6 ft) high and 5-10 m (16-30 ft) wide, with an erosional lower bounding surface (W cut-outs), and a gently sloping convex top distributed in wavy bedding. Wave ripples occur along the lower bounding surfaces and overlie the lenses as an X zone; typically, the ripples are continuous with those in the adjacent wavy bedding. The lenses become more numerous near the top, coalesce, and form tabular-bedded hummocky sequences. Bioturbation was not observed in this subfacies.

The tabular subfacies is 10-15 m (33-49 ft) thick and is composed of muddy, very fine to fine sand; grain size and sand content increase upward. The upper contact with the amalgamated subfacies is placed atop the TLfm. The tabular subfacies can be divided into upper and lower parts, both characterized by stacked, laterally continuous, layers. The major differences between the two parts are the scale and degree of development of the hummocky and wavy beds and the texture of the sediment. The contact between them, a hummocky scour, is marked by a change in colour from greenish grey (5 GY 6/1) below to light olive grey (5Y 6/1) above.

The lower portion of the subfacies is 1-2 m (3-7 ft) thick in the northern exposures and 4-5 m (13-16 ft) thick in the south. The sediment is composed of greenish grey, very fine, sandy mud to muddy, moderately sorted glauconitic sand (20 to 60 percent sand). The hummocky beds are 0.3 m (1 ft) thick; the W zones consist of single grey clay drapes, 1-2 cm (0.4-0.8 in) thick, atop thin X zones. The hummocky beds are intensely bioturbated with *O. nodosa* producing HbFbXbW sequences.

The upper portion of the subfacies consists of up to 8 m (26 ft) of light olive grey, muddy, moderately sorted, very fine to fine sand (50-80 percent sand). HFXW beds are 1.5-3 m (5-10 ft) thick. The L-sets of the hummocky zone average 1-2 m (3-7 ft) wide and 0.5 m (1.65 ft) thick. A fully preserved hummock-swale configuration 1 m (3.3 ft) thick and 10 m (33 ft) wide was observed at Site 1. Rare, isolated *O. nodosa* were observed in the muddy, very fine sand of the H zones. Well-defined zones of bioturbation are rare.

The amalgamated subfacies, composed of light olive grey, muddy, moderately sorted, fine sand (75-90 percent), is 7-11 m (23-36 ft) thick and typically weathers to moderate yellowish orange (10YR 6/6). The subfacies is characterized by Hb or W cut-out types of hummocky bedding, and the typical absence of W zones (F and X as well), probably due to lack of preservation, results in a massive sand unit. Bed contacts can be inferred from remnant W zones or the distribution of *O. borneenis* bioturbation zones. The individual L-sets are 0.3 m (1 ft) thick and 1-2 m (3-7 ft) wide; few complete L-sets were observed.

The stratigraphic position, colour, and lithology of the Timber Lake, as well as the presence of *Ophiomorpha*, are similar overall to the Timber Lake in the type area. However, they differ in sedimentary structures and petrology. The hummocky bedding common in the west is replaced by trough bedding in the type area, and the sediments of the study area lack the volcanic rock fragments noted by Waage (1968).

Colgate Member

This cross-bedded facies is composed of light olive grey (5Y 6/1), moderately to poorly sorted, muddy, very fine to medium sand (70-85 percent sand). The facies is 6-9 m (20-29 ft) thick, but may be as little as 2-4 m (7-13 ft) thick where Hell Creek channeling occurs. The basal contact is planar and erosional; the upper contact planar or channeled. Trough and epsilon cross-bedding are restricted to this facies.

At the majority of outcrops (2b, 2c, 6, and 7a), the facies is characterized by medium-to large-scale trough sets or cosets each overlain by a wavy layer; medium-scale tabular trough sets; and tabular, 0.3-0.5 m (1-1.7 ft) thick layers containing isolated scoop-shaped scours or tabular medium-to small-scale troughs in horizontal lamination, a type of scour and fill bedding. The scour and fill beds are capped by wavy layers, laminae discordances, or erosional ripples. The upper portion of the unit is composed of horizontal laminae with root molds.

At Site 4, the facies is characterized by large-scale trough cosets. The individual sets are 0.15-0.3 m (0.5-1 ft) thick and 2-3 m (7-10 ft) wide. In transverse section, the lower bounding surfaces are broadly curved and nonparallel; in longitudinal section, the sets have parallel to subparallel lower bounding surfaces and contain tangential laminae. Carbonaceous material, concentrated at the toes of forests, accentuates the lower bounding surfaces. A laterally continuous mudstone cobble lag occurs at the base, and rare, isolated *O. borneensis* occur in the trough sets.

At Site la, the facies features trough bedding at the base, overlain for over 100 m (326 ft) along the outcrop, by a 4 m (13 ft) thick set of epsilon cross-bedding. The foresets are accentuated by clay pebbles and carbonaceous drapes and are tangential with the erosional lower bounding surface.

At the type area south of Glendive, the Colgate Member is composed of multi-story, multilateral 6 m (20 ft) thick, upward fining sequences of lag to tabular-bedded (2 m [7 ft]), wedge-bedded (2 m [7 ft]), rippled lignite and shale (2 m [7 ft]) with carbonaceous material, and of pebble and cobble lags (Butler, 1980). In the study area, the combination of carbonaceous material in drapes and forsets and the occurrence of trough cross-bedding confirm these strata as the Colgate.

DEPOSITIONAL ENVIRONMENTS

BACKGROUND

The Fox Hills is Maastrichtian (latest Cretaceous) (Cobban and Reeside, 1952) and is diachronous (Waage, 1961, 1968; Gill and Cobban, 1973), being older in the southwest. Using ammonites, indexed with potassium-argon-dated betonites, Gill and Cobban (1973) demonstrated that the formation was deposited in North Dakota and eastern Montana 70.5 - 69 million years ago (Fig. 10). At that time, the climate was humid subtropical, prevailing winds were westerly, and seasonal changes were minimal (Dott and Batten, 1971). Fox Hills deposition occurred in a shallow marine environment associated with the Sheridan delta complex (Gill and Cobban, 1973).

The Sheridan Delta complex (*Baculities elasei* to *Baculities baculus* [71 to 70 mya]) rapidly (rate of 630 km/million years) advanced into the Williston Basin from the southwest and expanded northward along the border between Montana and North Dakota to separate the restricted Bearpaw deposition on the west from the open marine Pierre deposition on the east. Under this model, deposition would have occurred first in the study area followed by Areas 2 and 3. Crowder (1983) and Wheeler (1983) indicated that Area 1 may represent younger deposition associated with the filling of the Mosby Embayment from the west.

Subsequently, progradation rates diminished during the time of *Baculities baculus* to *Baculities clinobatus* (70 to 69 mya), and deposition shifted eastward, to the type area and other areas on the southeastern flank (Areas 5-7).

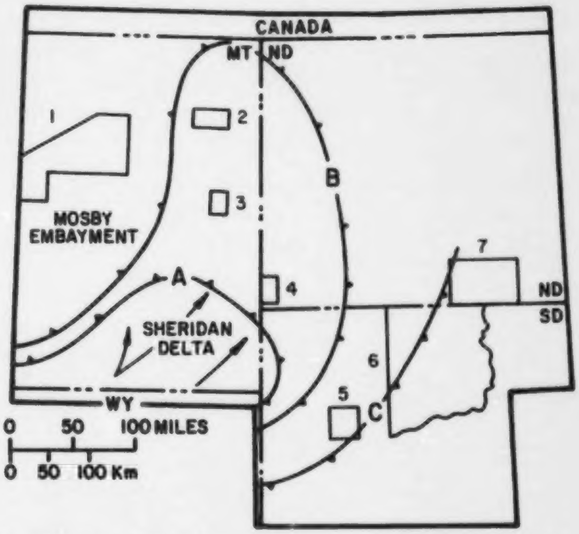

Figure 10: The distribution of strandlines through time for the Fox Hills regression (modified from Gill and Cobban, 1973). Strandline designations, corresponding to radiometrically indexed ammonite remains, are as follows: A=*Baculites baculus* (70.5-70 mya), B=*Baculites grandis* (70-69 mya), and C=*Baculites clinobatus* (69.5-69 mya).

STUDY AREA (AREA 4)

The section in Area 4 represents an upward shallowing marine sequence as evidenced by 1) a stratigraphic position between the underlying offshore marine sediments of the Pierre Formation and the overlying continental Hell Creek; 2) an upward increase in sorting and grain size from poorly sorted, sandy silt to moderately sorted, fine sand; 3) an upward change in trace fossils (all contained in hummocky bedding) from grazing (pasichnia) to dwelling (domichnia) structures; 4) an upward change in sedimentary structures from suspension dominated, through episodic, oscillation dominated, to current dominated; and 5) in the hummocky-bedded facies, an upward change from isolated hummocky beds to amalgamated beds.

Based on the interpretation of hummocky bedding as storm forms (*e.g.*, Dott and Bourgeois, 1982; Walder, 1984; Duke *et al.*, 1991) (Fig. 11), the Fox Hills sediments in the study area was laid down below storm wave base (Trail City Member), then formed storm-generated deposits on the inner shelf between storm wave base and fair weather wave base (Timber Lake Member), and finally shoreface, channel deltaic deposits (Colgate Member).

The Trail City deposit reflects two depositional regimes. The massive facies is a result of slow, steady sedimentation with which bioturbation kept pace to produce a largely reworked deposit. Remnant silt and clay interbeds exhibit diffuse, gradational contacts, indicating gradual changes in conditions and sediment types. Isolated hummocky beds resulted from episodic, storm events. Such sands were transported seaward and subsequently bioturbated during fair weather.

The predominance of sand, the dominance of episodic deposits, and the lack of directional current bedding in the hummocky facies of the Timber Lake Member suggest deposition episodically in the wake of storms on the inner shelf above the storm wave base, but below the normal fair weather wave base

(Dott and Bourgeois, 1982; Walker, 1984; Duke and others, 1991). Between depositional events, the substrate supported colonies of burrowing shrimp, producing *Ophiomorpha*. The upper portions of hummocky beds as well as the interevent wavy deposits were either not formed or not preserved in the higher-energy shoreward settings. Only zones of *O. borneensis*, a robust burrow form perhaps reflecting the higher-energy conditions, mark interevent horizons. The lack of preserved vertical burrow elements and the occurrence of burrow pellets along hummocky laminae suggest the periodic destruction of the upper portions of these burrow systems.

The Colgate strata (cross-bedded facies) are interpreted as predominantly shoreface (Walker, 1984, p. 148) to subaerial on the basis of: 1) upper flow regime bedforms including tabular or trough cosets, as well as scour fillings in horizontal lamination (upper plane bed); 2) the stratigraphic position between the hummocky-bedded facies and delta plain deposits as recognized in other examples from the rock record (*e.g.*, Harms *et al.*, 1975; Masters, 1967; Bourgeois, 1980; Dott and Bourgeois, 1982); 3) similar sedimentary structure sequence in modern settings (*e.g.*, Reineck and Singh, 1975; Howard and Reineck, 1981; Clifton *et al.*, 1971; Davidson-Arnott and Greenwood, 1976); and 4) a zone of root molds in the upper portion of the unit, suggesting shallow littoral or subaerial conditions. Abundant wavy layers indicate intermittent deposition, and the ubiquitous carbonaceous material and root molds, proximity to shore.

The brackish channel subenvironment, which corresponds to the entire member at Site 4, is characterized by an erosional basal contact, a basal lag (including, it appears, cobbles of the locally absent HCm), upward-fining texture, trough bedding, and *Ophiomorpha*. Epsilon bedding in the upper portion of the member at Site 1a may represent a tidal (?) channel point bar. At Site 6, the upper two-thirds of the Colgate is missing through local Hell Creek channeling.

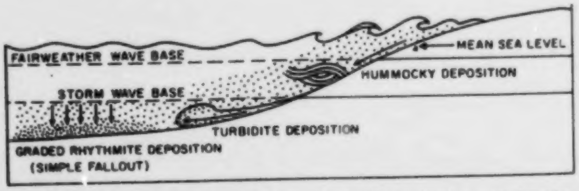

Figure 11: Model for the deposition of hummocky bedding and other storm deposits (from Dott, 1983).

REGIONAL RELATIONSHIPS

The Fox Hills Formation on the southeast flank has been formally subdivided since the 1950s, but terminology has not been systematically applied to outcrops on the north flank of the Blood Creek syncline (Area 1) and the west flank of the Poplar dome (Area 2). Reconnaissance and evaluation of published material suggest that the overall thickness, the sedimentary structure facies, and corresponding members, as defined in Area 4, are recognizable for the Fox Hills along the southwestern flank of the Williston Basin (Figs. 6, 7). The lower half of the Fairpoint Member in Area 5 contains strata similar to the Trail City and Timber Lake Members to the northwest.

Several authors have considered the Fox Hills along the southwestern rim as prodelta and delta front deposits (*e.g.*, Lepp, 1981; Wilde, 1984; Butler, 1980). Although sediment was undoubtedly stored along the coastline in deltas and Fox Hills strata may locally represent delta deposits, the Fox Hills Formation overall exhibits characteristics resulting from marine processes and is interpreted as representing environments marginal to or seaward of, the actively advancing delta systems. These observations generally agree with those of Crowder (1983), who described, however, Bullhead-like strata of deltaic origin in the upper Fox Hills on the southwestern flank of the Cedar Creek anticline. The upper Fox Hills represents tidal or deltaic channels as well as shoreface environments. In the vicinity of its type area, the Colgate is probably deltaic (highly meandering lower delta plain facies of Butler, 1980).

On the southeastern flank, in contrast, the lower Fox Hills is dominated by a major offshore bar complex corresponding to the sands of the Timber Lake Member. Waage (1968) indicated that this complex entered the type area from the northeast and grew to the south and that the abundant fauna fed on nutrients introduced by the sediment-bearing northerly longshore currents. The upper Fox Hills strata (Iron Lightning Member) are transitional from marine to continental and represent delta front, prodelta or lagoonal deposits (Bullhead), and channel or bar deposits (the Colgate) (Waage, 1968). The Linton Member has been described as estuarine (Klett and Erickson, 1976). Oyster reefs in the Colgate of the upper part of the Iron Lightning indicate shallow marine conditions.

The Meade County exposures (Area 5) show affinities to the southwestern flank and to the type area. The lower portion of the Fairpoint Member is similar to the offshore and inner shelf deposition of the southwestern rim. The overlying Stoneville Coal facies show coals alternating with *Ophiomorpha*-bearing marine or brackish sands. Marine to brackish conditions also characterize the White Owl Creek Member.

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CRETACEOUS SYSTEM STRATIGRAPHY AND SHALLOW GAS RESOURCES ON THE FORT PECK RESERVATION, NORTHEASTERN MONTANA

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ABSTRACT

Five shallow gas-bearing Cretaceous intervals have been identified on the Fort Peck Reservation of northeastern Montana: a relatively conventional gas reservoir in the Lower Judith River Sandstone, and shaly sandstones in the Gammon (Eagle equivalent), Niobrara (Martin Sandstone/ Medicine Hat equivalent), Greenhorn (Phillips Sandstone), and Mowry (Fish Scales Sandstone). Stratigraphic correlations have been carried from southwestern Saskatchewan through the Bowdoin Gas Field to the Reservation. Sparse, yet widely distributed gas shows confirm this relatively untested resource.

Each of these gas-bearing strata belongs to a recognized stratigraphic cycle characterized by thick shales overlain by progradational shaly sandstones and siltstones. The lowest cycle (Skull Creek to Mowry) contains considerable non-marine deposits, especially within the Muddy Sandstone which is thickly developed in the eastern part of the Reservation as a large valley fill network. The Reservation lies within transition zones for many of the facies associated with sea level changes along the western margin of the Cretaceous Seaway. Some individual sandstone units are not continuous across the Reservation. These and correlative units appear to be related to paleotectonic features defined by northwest-trending lineament zones and lineament zone intersections. Northeast-trending paleotectonic elements exert secondary influence on stratigraphic isopachs. Circular tectonic elements which carry through to Precambrian basement also have anomalous stratigraphic expression.

Conventional drilling has not been conducive to properly testing the Cretaceous gas potential on the Reservation, but by empirical well log analysis it is suggested that gas can be identified by various cross-over techniques. The Judith River yielded gas for field use at East Poplar.

STRATIGRAPHIC DISTRIBUTION OF *TITANOIDES* (MAMMALIA: PANTODONTA) IN THE FORT UNION GROUP (PALEOCENE) OF NORTH DAKOTA

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ABSTRACT

The type species of *Titanoides*, *T. primaevus*, is based upon a specimen discovered in 1913 northeast of Fort Union, near the Yellowstone-Missouri River confluence, in the upper Paleocene strata of the Fort Union Group in North Dakota. Although now known from numerous localities and represented by several species in the Rocky Mountain region, the stratigraphic and geographic context of the genoholotype has not yet been accurately reported. The relocation of the type locality of *Titanoides primaevus* and six other *Titanoides* occurrences in the state now allow these localities to be placed into both litho- and biostratigraphic context. The lowest record is the middle Tiffanian age (late Paleocene) Judson Locality, which is near the Slope-Bullion Creek formational contact, occurring about 79 ft (24 m) above the Cannonball Formation. The remaining localities are all in the lower portion (within 164 ft [50 m] of the base) of the Sentinel Butte Formation. The type locality, named the Witter Locality, is approximately 72 ft (21.9 m) above a recently discovered locality that has produced mammals of middle Tiffanian age. This may indicate that the Witter Locality is slightly younger than the Judson Locality, but both localities may also represent the same biochron. The purportedly youngest occurrence in the state is the Riverdale Locality, which was previously reported from the Tongue River Formation, but has since been reinterpreted as the Sentinel Butte Formation. The mammalian fauna from the Riverdale Locality is from a stratigraphically higher horizon than the Judson Locality, but is still middle Tiffanian. Although known from younger localities elsewhere (late Tiffanian and possibly early Clarkforkian, representing latest Paleocene), all of the known *Titanoides* localities in North Dakota are middle Tiffanian.

INTRODUCTION

TITANOIDES

Titanoides is a member of the Order Pantodonta, comprised of small to large herbivores, which occur in the Paleogene of Europe, Asia, and North and South America. In North America, the order is found in middle Paleocene to Early Eocene faunas. The type locality of the genoholotype of *Titanoides* is from the Witter Locality in North Dakota. *Titanoides* is distinct from most other pantodonts in having claws on the forefoot (the nature of the hind foot is unknown) rather than the usual hooves. On the basis of certain skeletal features, Patterson (1939) was the first to suggest that the claws were used for digging. This feature, together with the development of the cervical vertebrae and the nature of the sabrelike upper canines and hooklike lower canines implied that *Titanoides* fed on roots and tubers (Patterson, 1939; Coombs, 1983). *Titanoides* has the distinction of being among the largest mammals of its day, approximately three feet tall (one meter) at the shoulder and probably weighing in the vicinity of 200 pounds (100 kg). The pantodonts gave rise to no surviving lineages. The ecological niche they filled was gradually taken over by the modern ungulate groups beginning in the Eocene (*e.g.*, Perissodactyla).

PURPOSE OF CURRENT RESEARCH

This paper examines the geologic record of *Titanoides* in North Dakota. At present, this includes seven localities, which range in representation from single poorly preserved specimens at some to exquisitely preserved skulls at others. Even though the published history of these occurrences dates back to 1917, the previously documented knowledge of the localities is vague. In most cases, both geographic location and stratigraphic horizon of the fossil locality have been imprecisely or incorrectly reported. The purpose of this paper is to summarize present research, in order to provide the geographic and stratigraphic context of *Titanoides* in North Dakota for more extensive taxonomic and stratigraphic studies.

STRATIGRAPHY

The record of *Titanoides* is confined to Tiffanian-age strata in North Dakota. These strata form the widespread Fort Union Group of the Williston Basin (Figs. 1 and 2). In ascending stratigraphic order, *Titanoides* localities occur in the Slope, Bullion Creek, and Sentinel Butte Formations. The Slope and Bullion Creek Formations were recently defined by Clayton *et al.* (1977) and represent, as they concern this study, the Tongue River Formation of previous usage. The use of the names Slope and Bullion Creek has largely been ignored by the U.S. Geological Survey, and Bullion Creek is not mapped as such by Carlson (1982, Grant and Sioux Counties; 1983, Morton County; 1985, McKenzie County) for the North Dakota Geological Survey. Correlation of the Slope Formation with strata previously assigned to basal Tongue River strata in south-central North Dakota will require more detailed mapping to justify its use on the basis of the criteria upon which it was introduced. As the main criterion, the white siliceous bed in Grant and Morton Counties must be shown to be stratigraphically equivalent to the lithologically similar Rhame bed (Wehrfritz, 1978), used to mark the top of the Slope-Bullion Creek contact in Slope, Golden Valley, and Billings Counties (Moore, 1976; Clayton *et al.*, 1977). This white marker bed was thought to represent a widespread unconformity (Clayton *et al.*, 1977; Carlson, 1983) that would provide an unambiguous stratigraphic marker. The Slope Formation, however, has been shown to be difficult to map east of the Little Missouri River (Carlson, 1982, 1983). Ongoing biostratigraphic studies indicate that the contact between the Slope and Bullion Creek Formations may be significantly diachronous (Hartman and Kihm, 1991), which suggests, in the absence of other evidence, a lack of equivalence between the white beds of central and western North Dakota. With these reservations noted, this paper will employ the formation names Slope and Bullion Creek.

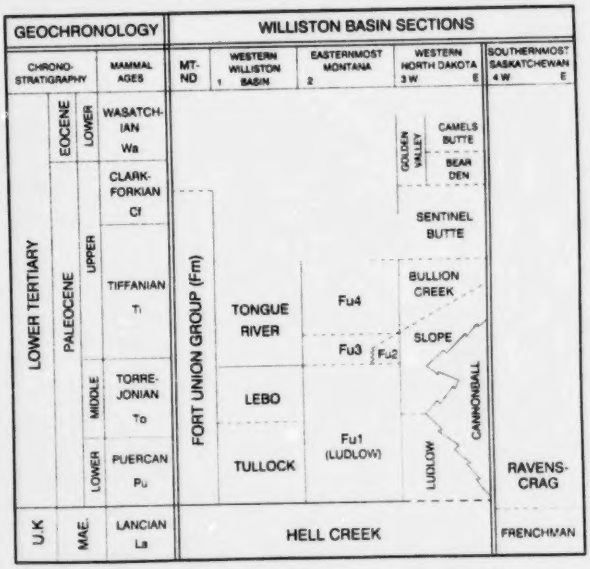

Figure 1: Chronostratigraphic correlation of Paleocene lithostratigraphic units in the Williston Basin

TITANOIDES IN NORTH DAKOTA

The published account of *Titanoides* in North Dakota (e.g., Archibald *et al.*, 1987) has perpetuated, for the lack of more precise knowledge, inaccuracies about the location and stratigraphy of its occurrence. The discovery of the most important specimens was largely accomplished by local ranchers and townspeople, or by professional naturalists who simply collected their finds for the record. The following account provides both the historical basis and our current knowledge about the geographic location and stratigraphic position of each locality. The localities are presented in numerical order (Fig. 2).

Witter Locality (L5608)

The first specimen of a new species of extinct mammal to be discovered in North Dakota was collected by Vernon Bailey in July of 1913 as part of a biological survey program (which later became the U.S. Fish and Wildlife Service) conducted by the U.S. Department of Agriculture. The partial lower jaw was described and assigned to a new genus and species, *Titanoides primaevus*, by Gidley (1917). Gidley (p. 432) reported the find as from "about 3 miles northeast of Buford (Fort Union of early days), North Dakota, Fort Union formation (type section)" (Fig. 2) (Buford and Fort Union do not represent the same location). The locality has subsequently been referred to by some as the "Buford" locality. As explained by Simons (1960), the "x" on the photograph, intended to indicate the location of the specimen of *T. primaevus*, and the photograph published by Gidley (1917, pl. 36) were errors. Bailey subsequently provided additional photographs to Glenn Jepsen of Princeton University, who sent Robert Witter and Frank Goto to North Dakota in 1951 to relocate the locality. Neither the locality nor Jepsen's specimens were relocated as reported by Holtzman (1978). Reference to the locality in this report as the Witter Locality reflects our specific knowledge as to the site collected by Witter and Goto, and its use permits no confusion with the incorrect location illustrated by Gidley (1917). As of yet, the photographs used by Witter to document the equivalence of his relocation with the original Bailey site have not been found. In July of 1990, using photographs provided by Robert Witter, the type locality of *Titanoides primaevus* was relocated (Fig. 3). Preliminary investigations found additional bone in the minor quarry excavated by Witter and Goto. New field studies will be undertaken to determine the potential for additional type material.

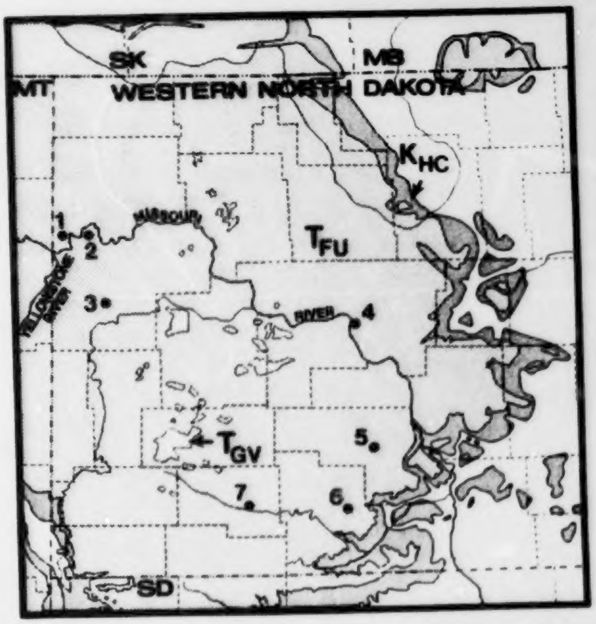

Figure 2: *Titanoides* localities in the North Dakota. KHC = Hell Creek Formation. TFU = Fort Union Group. TGV = Golden Valley Formation. Explanation of locality numbers 1-7: 1 = Witter Locality; 2 = White's River Basin Survey Site 3; 3 = Cross Locality; 4 = Riverdale Locality; 5 = Judson Locality; 6 = Brisbane Locality; and 7 = Mott Dump Site.

The stratigraphic horizon of the Witter Locality was first reported as being from the type section of the Fort Union Formation (Gidley, 1917) (note that no specific type section for the Fort Union Group of Meek and Hayden, 1862, exists). W.T. Thom, Jr., in a personal communication to Jepsen (1940, p. 242), stated that the stratigraphic level of the specimen found near Buford may be in either the Tongue River or lower Sentinel Butte Formation. Simons (1960, p. 33) reported the type locality as from the "Sentinel Butte shale member" of the "Fort Union formation." This placement was confirmed by Brown (1962), who stated that *Titanoides primaevus* was found in Meek and Hayden's Fort Union "type section," in a part laterally equivalent to the Sentinel Butte Member. Sloan (1969) subsequently illustrated the "Buford" locality as about 300 ft (91 m) above the base of the Sentinel

Butte Formation. Archibald *et al.* (1987, fig. 3.2, locality 91, part 1) illustrated the type locality as occurring at or near the Tongue River-Sentinel Butte Formation contact (see Fig. 4).

Our knowledge of the location of the type locality of *Titanoides primaevus* is based on two collecting trips: the original discovery by Bailey and its relocation by Witter. Using photographs taken by Witter, the collections of *Titanoides primaevus* are from the NE¼ NW¼ SW¼ SW¼ SE¼ Sec. 32, T. 153 N., R. 103 W., Trenton SW Quadrangle (1968), Williams County, North Dakota. The locality is at an elevation of about 2240 ft (682.8 m) on a sloping bench of a minor south-directed promontory. The stratigraphic horizon is interpreted to be about 157 ft (47.9 m) above the contact of the Bullion Creek and Sentinel Butte Formations as interpreted from two lines of evidence. The first is the relative difference (along strike) in elevation between the locality and the contact between these formations, believed to be poorly exposed one mile to the south (just north of the Burlington Northern tracks). The second determination was made from a three-point solution from the area of Snowden Siding in T. 26-27 N., R. 59 E., Roosevelt County, Montana, where the formational contact is relatively well exposed and delimited, to the Hardscrabble Quarry section in Sec. 27, T. 153 N., R. 103 W., which is believed to include the basal sandstone of the Sentinel Butte Formation. The two methods produce essentially identical results.

Figure 3: The type locality of *Titanoides primaevus* (Witter Locality). Easterly view of the type locality of *Titanoides primaevus* (Witter Locality) in the lower portion of the Sentinel Butte Formation. Note presence of "star" along trend of arrow.

White's River Basin Survey Site 3 (L5383)

In August of 1951, T.E. White, along with Easton and Harrup, discovered and collected a skull White identified as *Pantolambda*. The original reference to this collection (other than White's field notes, which have not, as yet, been found by the authors) was on a form of the Garrison Reservoir "River Basin Surveys" conducted by the Smithsonian Institution (White, 1951). This specimen was incorrectly reported, based on what we believe to be a map-reading error, as coming from the "NW¼ Sec. 29, T. 153 N., R. 102 W., McKenzie County, North Dakota." This legal location, which is actually in the Missouri River (Hartman, 1984), has been cited by subsequent authors. The first published report by Gazin (1956) unequivocally assigned the specimen from this locality to *Titanoides primaevus*. Simons (1960), reporting the original incorrect legal location from a museum label, published this record of *T. primaevus* in his monograph on the pantodonts. Simons' identification and location were included by Holtzman (1978) in his study of the Paleocene mammals of North Dakota. In keeping with the original legal location, Archibald *et al.* (1987; fig. 3.2, locality 91, part 3) reported the locality to be Williams County, North Dakota.

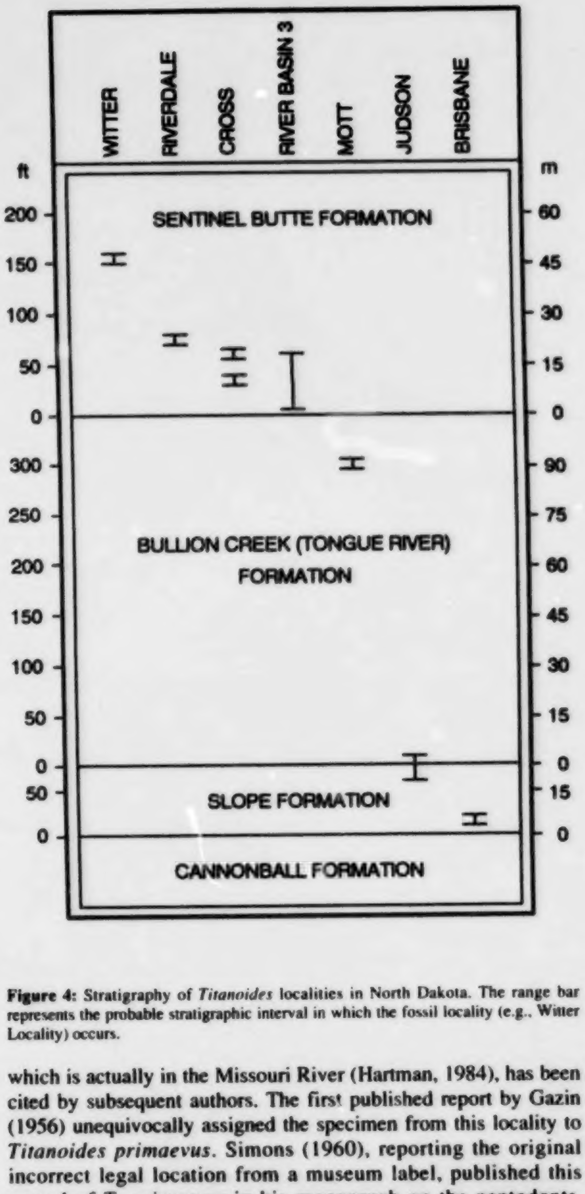

Figure 4: Stratigraphy of *Titanoides* localities in North Dakota. The range bar represents the probable stratigraphic interval in which the fossil locality (e.g., Witter Locality) occurs.

White (1951) also reported his *Titanoides* to have come from the Tongue River Member of the Fort Union Formation (see Fig. 4). Simons (1960) cited the occurrence as in the Sentinel Butte Shale of the Fort Union Formation. Holtzman (1978) suggested that the specimen was probably from Tongue River strata by inspection of the Bedrock Geologic Map of North Dakota by Carlson (1969; 1:1,000,000). As noted above, Archibald *et al.* (1987) illustrated all of the records of *Titanoides* in far-western North Dakota as near or at the contact of the Tongue River and

Sentinel Butte Formation.

The location and stratigraphic position of T.E. White's (*op. cit.*) River Basin Survey Site 3 are not known precisely, but a placement can be made on the basis of present field studies and existing unpublished and published data. White (1951; U.S. National Museum specimen label) reported his *Titanoides* locality (site no. 32 MZ 201) as from the south side of the Missouri River, 2.5 to 3 mi (4 to 4.8 km) west of U.S. 85 (see also an unpublished Missouri Basin Project report by Metcalf *et al.*, 1953). The locality is herein interpreted to be in the NW¼ Sec. 29, T. 153 N., R. 101 W., Williston SW Quadrangle (1979), McKenzie County, North Dakota (Fig. 2), at an elevation of about 1885 ft (574.6 m) at the base of north-facing bluffs overlooking the Missouri River. The stratigraphic horizon is interpreted to be between 5 and 62 ft (1.5 to 18.9 m) above the Bullion Creek-Sentinel Butte contact. This stratigraphic range is based on the correlation of the stratigraphic interval by Freers (1970, pl. 3; see also Clayton, 1980) through his measured sections 2 through 6. White (1951) provided the following information about the occurrence of *Titanoides* on his River Basin Surveys paleontology form (note that White reported measuring a section through this interval, but this information has not been found): "Bone occurs in buff sandy clay with limonite knodules [sic] about 20 feet [6.1 m] below a 1 ft [0.305 m] coal seam and just above a 24 inch [0.610 m] layer of blue gray shaley clay." This interval has been relocated and corresponds to Freers' measured section 5 (Freers, 1970, pl. 3).

Cross Locality (L5377a and b)

Three individuals of *Titanoides* were incorrectly reported from the NW¼ Sec. 30, T. 148 N., R. 101 W. (see revision below), McKenzie County, North Dakota (Holtzman, 1978; McKenna, pers. comm., 1988; McKenna, pers. comm., 1991, stated that possibly only two individuals are represented) (Fig. 2). This locality, referred to by Archibald *et al.* (1987: fig. 3.2, locality 91, part 2) as "McKenna's *Titanoides* material, McKenzie Co.", was grouped with the other "miscellaneous" *Titanoides* from North Dakota as occurring at or near the Tongue River-Sentinel Butte Formation contact. "McKenna's *Titanoides*," here referred to as the Cross Locality, was discovered by Morris and Ralph Cross, of Alexander, North Dakota, and shown to Malcolm and Priscilla McKenna in July of 1963. Through the course of their field examination, two mammal-bearing sites were collected, one on either side of a valley, referred to as the "north" (L5377b) and "south" (L5377a) sites by McKenna, the south being the main site (McKenna, pers. comm., 1988). The material from these two sites has yet to be reexamined.

With the assistance of Morris and Ralph Cross, the Cross Locality was relocated and mapped in August, 1990. Collections were made at the "south site" (L5377a), which is located in northwest-facing exposures near the center of the north edge of the SW¼ SW¼ SW¼ Sec. 19, T. 148 N., R. 101 W., Red Wing Creek Quadrangle (1972), McKenzie County, North Dakota (Fig. 2). The elevation of the collecting horizon is placed at about 2293 ft (698.9 m). The "north site" (L5377b) is believed to be located in the SE¼ NE¼ NW¼ SW¼ SW¼ of Sec. 19. The elevation of this site is estimated at 2265 ft (690.4 m). The Cross Locality is some distance from exposures of the Bullion Creek-Sentinel Butte contact. The locality was placed relative to this contact through an approximate correlation of the bentonitic ash known as the "blue bed" and its position relative to the contact. The blue bed was mapped at a number of locations to the north of the Cross Locality and was found to occur at a relatively consistent elevation of 2390 ft (728.5 m). Thus the south site occurs about 97 ft (29.6 m) below the blue bed. The stratigraphic horizon of the blue bed is about 159 ft (48.5 m) above the Bullion Creek- Sentinel Butte contact. This value was derived utilizing the correlations of various beds, including the blue bed, in the Sheep Butte and Bear Butte (not Bare Buttes of Fisher) sections measured by Fisher (1953). Thus, the south and north sites of the Cross Locality occur 62 ft (18.9 m) and 34 ft (10.4 m), respectively, above the base of the Sentinel Butte Formation (Fig. 4).

Riverdale Locality (L1)

The Riverdale Locality was discovered in 1959 or 1960 and initially collected from 1960 to 1963 by Bernice Vinje and her family of Hazen, North Dakota (Vinje, pers. comm., 1990). In June of 1962, William Turnbull, with the Vinjes, screened and picked additional vertebrate and molluscan specimens (Turnbull, 1962; Holtzman, 1978; Hartman, 1976, 1984). In 1964, the locality, which has produced the best preserved specimens of *Titanoides primaevus*, was lost to the rising waters of the Garrison Reservoir, otherwise known as Lake Sakakawea. The variety of taxa to be described from the Riverdale Locality noted by Turnbull (1962), was studied by Holtzman (1978). A Ti4? age was subsequently assigned to the Riverdale Locality by Archibald *et al.* (1987) in their review of North American Paleocene mammalian biochronology. Holtzman reported the locality as from "approximately Sec 22, T147N, R84W, McLean County, North Dakota" (see revision below). This location was based on collection data recorded and subsequently revised by Turnbull without the aid of maps or field checking (Turnbull, written comm., 1990).

Holtzman (1978, p. 12) reported the Riverdale Locality as occurring "within the upper 30 meters [98.4 ft] of the Tongue River Formation" (Fig. 4). On the basis of geologic mapping by Bluemle (1971) and Carlson (1973), Holtzman (1978, p. 12) stated: "Although there is little doubt that the stratigraphic position of this locality is close to the contact between the Tongue River and Sentinel Butte Formations, its precise horizon is uncertain." Their estimate is about 200 ft (61 m) above the contact as interpreted by Groenewold *et al.* (1979). Groenewold's placement is consistent with the air photo interpretation and mapping that resulted in the State Geologic Map of North Dakota (Clayton, 1980). The Groenewold *et al.* (1979) Bullion Creek-Sentinel Butte contact is also consistent with the interpreted relative position of the contact with the underlying Tavis Creek coal bed (Carlson, 1983). Assignment of the strata above the thick coal (Tavis Creek) at the base of the Garrison Dam spillway to the Sentinel Butte Formation is in agreement with the coal correlations by Hartman (1984, pl. 5, cross section K) and our more recent field observations.

With the drought of the late 1980s and early '90s, the level of Lake Sakakawea has fallen sufficiently that a large diameter pipe, known relative to the original Riverdale Locality, is again visible. On a North Dakota Paleontological Society field trip in October of 1989, Mrs. Vinje noted that a mollusk-rich lithology (L5507) similar to that which produced the Riverdale specimens had been exhumed (Campbell, pers. comm., 1990). This knowledge prompted field studies of the area and discussions on the history of the Riverdale Locality with Mrs. Vinje.

The original Riverdale Locality has probably been lost to erosion. New field observations, however, estimate the locality to have been about the center S½ NW¼ Sec. 34, T. 147 N., R. 84 W. This location is about two miles (3.2 km) south of the legal description referenced by Holtzman (1978). The elevation of the Riverdale Locality was given by Holtzman (1978) as 549 m (1801.2 ft), which was determined from water level records provided by the Army Corps of Engineers. The Riverdale Locality is estimated to be about 76 ft (23.2 m) above the base of the

Sentinel Butte Formation, as interpreted by Groenewold *et al.* (1979). This level was determined by correlating measured and drill hole sections from north of Government Bay to south of Garrison Dam. The Riverdale Locality is approximately 80 ft (24 m) above the Tavis Creek coal bed (of Groenewold *et al.*, 1979) and 82 ft (25 m) below the Hagel coal split exposed in the adjacent west-facing cliff.

Judson Locality (L6 and L8)

The Judson Locality (Holtzman, 1976, 1978) consisted of four distinct sites probably in the same lithologic unit (Hartman, 1984). These sites (numbers 1-4) were treated faunally as one locality by Holtzman (1976, 1978) but as separate localities by Hartman (1976, 1984). The original Judson sites (1 and 2) were discovered in 1962 by Joseph and Gordon Bell of Bismarck, North Dakota. The subsequent collections made (between 1963-1973) from these sites by Steve Lund were treated by him as one collection. Thus a very worn (possibly deciduous) P^4 (SMM P77.6.81), questionably assigned to *Titanoides primaevus* by Holtzman (1978), could be from either Site 1 (L8) or Site 2 (L6). The better preserved but fragmentary lingual left P^4 (SMM P77.6.80) was collected by Holtzman in 1974 at Site 2 (L6) (Holtzman, pers. comm., 1991).Sites 1 and 2 of the Judson Locality were covered and graded by the Morton County Highway Department in 1973 (Holtzman, 1978). Site 2 is at the west edge of the NW¼ SW¼ SW¼ sec. 27, T. 138 N., R. 84 W., Green Ridge Quadrangle (1980), Morton County, North Dakota. The elevation of Site 2 was determined by Holtzman (1976) from an altimeter survey as 638 m (2093.2 ft). This value may be slightly high by about 10 ft (3 m) as topographically interpreted from the Green Ridge Quadrangle. Site 1 was located on the opposite side of the section line road on the east edge of the SE¼ NE¼ NE¼ Sec. 28, T. 138 N., R. 84 W. The elevation of Site 1 was determined by Holtzman (1976) from an altimeter survey as 637 m (2089.9 ft), which is consistent with the Green Ridge Quadrangle.

Holtzman (1978) estimated that the Judson sites were 24 m (78.7 ft) above the contact between the Tongue River and underlying Cannonball Formations. This estimate is consistent with that given by Hartman (1984) and with an elevation for the top of the Cannonball Formation derived from Carlson (1983). The precise placement of Judson Sites 1 and 2 relative to the Slope-Bullion Creek contact is not determined, but is believed to be near the contact. Hartman (1984) assigned the Judson Localities to the Slope Formation. Unpublished observations on the "white bed" (Carlson, 1982), a few miles southwest of the localities, place it at about the elevation of the Judson Locality.

Brisbane Locality (L5385)

The Brisbane Locality was discovered by Richard Holtzman in 1973 (Holtzman, 1976, 1978). Holtzman (1978, p. 13) stated that because of the relatively high concentration of fossils from Brisbane, collection at the Judson sites was discontinued. Holtzman reported that the 4.6 metric tons (5.1 tons) screened from the Brisbane Locality during 1973 and 1974 produced 310 identifiable specimens. Of these, only one, a poor tooth fragment, was reported as "probably" *Titanoides primaevus*, "but more certainly represents some very large pantodont" (Holtzman, 1978, p. 61). On the basis of present studies, the specimen is questionably assigned to *Titanoides.*

The Brisbane Locality was a small low-lying roadcut exposure that has since been graded and grassed by the Grant County Highway Department (Holtzman, pers. comm., 1991). Vertebrates were collected by Holtzman (1978) from west-facing exposures on the west edge of the NW¼ NW¼ SW¼ SW¼ Sec. 3, T. 133 N., R. 86 W., Brisbane Quadrangle (1980), Grant County, North Dakota. The elevation of the locality was given by Holtzman (1978) as 723 m (2372.0 ft), which compares well with that determined from the Brisbane Quadrangle. The Brisbane Locality was placed by Holtzman (1978) at about 7 m (23.0 ft) above the Cannonball Formation, on the basis of mapping by Lloyd (1914). The Cannonball-Tongue River contact as mapped by Carlson (1982) indicated the locality to lie within 12 to 23 ft (3.7 to 7.0 m) of the base of the Tongue River Formation (of Carlson's usage), i.e., within the Slope Formation, as based on an estimated thickness of 80 ft (24.4 m) for the latter formation in this general area (Carlson, 1982, pl. 3).

Mott Dump Site (L4954)

Vertebrates, believed to be derived from a mollusk-bearing bed, were first collected by John Hoganson on a spoil pile at the Mott Sanitary Landfill in June of 1986. To date, only a single tooth, representing an undetermined species of *Titanoides*, has been found. No other research has previously been undertaken concerning the specimen.

The fossil-bearing exposure, in a relatively new east-west trending trench at the Mott Sanitary Landfill, is in the NW¼ NE¼ SE¼ SE¼ Sec. 36, T. 134 N., R. 93 W., Mott South Quadrangle (1966), Hettinger County, North Dakota. Elevation of the producing horizon is about 2390 ft (728.5 m). The Mott Dump site is estimated to be about 50 ft (15.2 m) below the top of the Bullion Creek Formation. This value was derived from a transfer of the mapped Tongue River-Sentinel Butte formational contact by Trapp and Croft (1975) to topographic maps and the determination of the relative difference in elevations between the contact and the site. The estimated stratigraphic position is considered to be relatively accurate in that 1) the strata between the dump and the contact, a distance of about 1 mile (1.6 km), are essentially horizontal; 2) there is a close correspondence between the mapped contact and the topography; and 3) the contour interval of 10 ft (3.0 m) permits a relatively precise interpretation of elevation.

CHRONOSTRATIGRAPHY OF LOCALITIES

MAMMALIAN CONTEXT

The use of the evolutionary record of mammals in North America to define time units (biochrons) has been very successful for nonmarine Tertiary strata. The first formal designation of mammalian biochrons in North America was by Wood *et al.* (1941), who divided each of the Tertiary epochs into three to five "provincial land mammal ages." Considerable modification and refinement has taken place since 1941, but the basic framework remains. The most recent revision of Paleocene mammalian biochronology was by Archibald *et al.* (1987) and is, in part, followed here, with important modifications regarding the placement of the Paleocene-Eocene boundary from Gingerich (1989a, b) (Fig. 5.).

There are four land mammal ages within the North American Paleocene record (Fig. 5). The oldest is the Puercan, succeeded by the Torrejonian, Tiffanian, and Clarkforkian. The record of *Titanoides* in North Dakota is confined to the middle Tiffanian. The Tiffanian is defined as the time between the first appearance datum (FAD) of *Plesiadapis* (a primitive primate-like mammal) and the first appearance of the Order Rodentia. In turn, it can be divided into six numbered biochrons, Ti1 through Ti6. These are defined by the FAD's of species of *Plesiadapis*, except for Ti6, which is defined as the time between the FAD of *P. gingerichi* and the Order Rodentia (Fig. 5). Besides *Plesiadapis*, each numbered subdivision is characterized by a number of species making their FAD, their last appearance (LAD), and by index fossils.

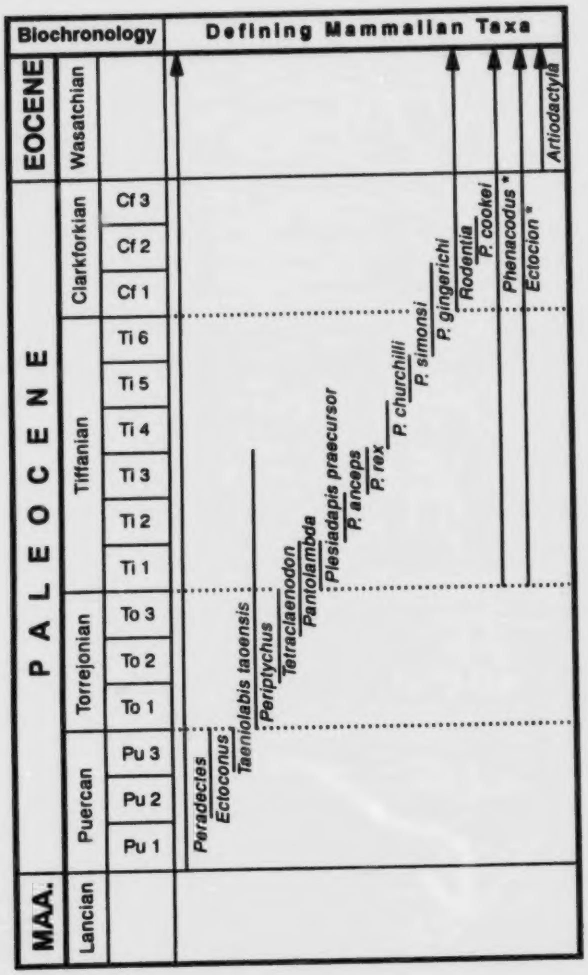

Figure 5: Paleocene mammalian biochronology - Definition of the mammal zones (as given on figure). Mammalian biochronologic framework from Archibald *et al.* (1987) and Gingerich (1989a, b). The "*" indicates that the Cf3 biochron is defined on the basis of *Phenacodus-Ectocion* acme zone.

THE AGE OF *TITANOIDES* IN DAKOTA

The North Dakota Paleocene record of *Titanoides* occurs in the Slope, Bullion Creek, and Sentinel Butte Formations of the Fort Union Group. The stratigraphically lowest locality is Brisbane from the Slope Formation, or lowermost Tongue River Formation of previous usage (Hartman, 1984). This locality has produced a tooth fragment that Holtzman (1978) tentatively suggested could represent *Titanoides*. The Brisbane Locality has produced a moderate-sized mammalian fauna identifiable as Ti3 in age (Holtzman, 1978; Archibald *et al.*, 1987). Above Brisbane, near the Slope-Bullion Creek formational contact, are the Judson sites, which have produced only a marginally better pantodont record, consisting of an isolated worn deciduous? tooth and a tooth fragment. These specimens were referred to *Titanoides primaevus* by Holtzman (1978). The Judson Locality has produced a well-documented Ti3 fauna (Holtzman, 1978; Archibald *et al.*, 1987). Above the Judson Locality, in the uppermost part of the Bullion Creek Formation, is the Mott Dump Site, which has produced a single lower premolar that can be tentatively identified as *Titanoides*. No other mammals are yet known from this site.

The stratigraphically lowest well-preserved specimen of *Titanoides primaevus* is from White's River Basin Survey Site 3, from strata low in the Sentinel Butte Formation (Fig. 4). This site has produced a skull but no other mammals. The Cross and Riverdale Localities are near the same horizon in the Sentinel Butte Formation. The Cross Locality has produced fragmentary dentitions of three individuals tentatively identified as *Titanoides* (Holtzman, 1978; McKenna, pers. comm., 1988), but no other mammals. The Riverdale Locality has produced the best specimen of *Titanoides*, consisting of a well-preserved skull with some skeletal material and an isolated tooth referable to *T. primaevus*. The Riverdale Locality has also produced enough mammalian material to suggest an age younger than the Judson Locality, possibly Ti4 (Holtzman, 1978; Archibald *et al.*, 1987). Although the locality is significantly higher in the section than the Judson Locality (of a Ti3 age), a Ti4 age cannot currently be assigned to the Riverdale Locality with confidence. The Witter Locality has produced the stratigraphically highest occurrence of *Titanoides* in North Dakota and represents the type locality of *Titanoides primaevus*. The Witter Locality collection consists of both lower and upper dentitions, and fragments of a skull. Additional bone is still in place at the Witter Locality and may represent postcranial remains of the type specimen. The age of the locality is still open to question, but on the basis of preliminary identifications of mammalian material from a nearby locality (L5500b), it is Ti3, or possibly Ti4.

The most recent revision of *Titanoides* recognized five species with the oldest being *T. simpsoni* from the Torrejonian of the Crazy Mountains Basin, Montana (Simons, 1960). Gingerich and Childress (1983; see also Lucas, 1984) synonymized this species with *Pantolambda intermedium*, thereby restricting *Titanoides* to Tiffanian and younger deposits.

Outside of North Dakota, *Titanoides primaevus* is known from Wyoming and Colorado. The Wyoming record comprises two occurrences: 1) the *Titanoides* locality in the Fort Union Formation of Bison Basin, Fremont County (Gazin, 1956); and 2) from the vicinity of Princeton Quarry in the Polecat Bench (Fort Union of others) Formation of the Clark's Fork Basin, Park County (Rose, 1981) (Table 1). Gingerich and Childress (1983) suggested that the specimen from the Polecat Bench Formation may represent *T. major* [sic] (actually *T. majus*). This occurrence was also suggested to be below the level of the Princeton Quarry and probably Ti2 in age (Gingerich, pers. comm., 1991). This leaves the Bison Basin specimen as the only documented Ti5-age record of *Titanoides primaevus* (Fig. 5). No specimens of *T. primaevus* are known from Ti6-age localities, but several specimens have been reported from the early Clarkforkian (Cf1) Plateau Valley local fauna of Mesa County, Colorado (Simons, 1960; Archibald *et al.*, 1987). Patterson (1939) originally identified these specimens as *Sparactolambda looki*. Gazin (1956) suggested that this species was synonymous with *Titanoides primaevus*, a reassignment followed by Simons (1960). Reconsideration of this synonymy is warranted in that the Clarkforkian material is quite possibly distinct from the holotype (Kihm, unpub.). If the Plateau Valley specimens represent a

Table 1: The Record of *Titanoides primaevus* and of *Titanoides* specimens from North Dakota

PALEOCENE

CLARKFORKIAN (Cf1)
- Plateau Valley local fauna (Mesa Co., northwestern Colorado)
  - Debeque Formation ("Wasatch" Formation)
    - "***Titanoides primaevus***" (substantial record); probably referable to a taxon other than ***T. primaevus***
      - Patterson (1939), Gazin (1956), Simons (1960), Kihm (1984), this study

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TIFFANIAN (Ti6) no record

TIFFANIAN (Ti5)
- Bison Basin ***Titanoides*** Locality (Fremont Co., south-central Wyoming)
  - Fort Union Formation
    - ***Titanoides primaevus*** (one maxilla, consisting of three teeth)
      - Gazin (1956)
- vicinity of Princeton Quarry (Park Co., northwestern Wyoming)
  - Polecat Bench Formation (Fort Union Formation)
    - "***Titanoides primaevus***" (partial dentition of one individual)
      - Simons (1960), Rose (1981); this specimen was suggested to be referrable to *T.* ***majus*** (Gingerich and Childress, 1983); its locality is significantly below Princeton Quarry and is probably Ti2 in age (Gingerich, verbal communication, 1991)

TIFFANIAN (?Ti3-Ti4)
- Type-Locality (Williams Co., northwestern North Dakota)
  - Sentinel Butte Formation
    - ***Titanoides primaevus*** (lower jaw and fragmentary skull with dentition; post-cranial material under investigation)
      - Gidley (1917), Simons (1960), Gazin (1956), Holtzman (1978), this study
- Riverdale Locality (McLean Co., north-central North Dakota)
  - Sentinel Butte Formation
    - ***Titanoides primaevus*** (well-preserved skull with some postcranial material and one isolated tooth)
      - Holtzman (1978), this study (see text for discussion of age)

TIFFANIAN (CF. Ti3)
- Cross Locality (McKenzie Co., northwestern North Dakota)
  - Sentinel Butte Formation
    - ***Titanoides*** sp. (fragmentary dentitions from three* individuals)
      - Holtzman (1978); * see text for discussion
- White's River Basin Survey Site 3 (McKenzie Co., northwestern North Dakota)
  - Sentinel Butte Formation
    - ***Titanoides primaevus*** (skull)
      - Gazin (1956), Simons (1960), Holtzman (1978)
- Mott Dump Site (Hettinger Co., southwestern North Dakota)
  - Bullion Creek Formation
    - ***Titanoides*** sp. (single tooth)
      - this study

TIFFANIAN (Ti3)
- Judson Locality (Morton Co., south-central North Dakota)
  - Bullion Creek Formation
    - ***Titanoides primaevus***? (two isolated teeth)
      - Holtzman (1978), this study
- Brisbane Locality (Grant Co., south-central North Dakota)
  - Bullion Creek Formation
    - ***Titanoides***? sp. (poor single-tooth fragment)
      - Holtzman (1978), this study

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different taxon, *Titanoides primaevus* may be restricted to the Tiffanian.

DISCUSSION

The main biochronologic summaries involving Paleocene mammals from North Dakota were published by Sloan (1969, 1987) and Archibald *et al.* (1987). Sloan (1969) placed the Buford Locality (herein referred to as the Witter Locality) late in the Clarkforkian. This age assessment was probably based inferentially on Sloan's rather high estimate of the locality's horizon in the Sentinel Butte Formation. Sloan (1987) subsequently revised his age of the "Buford" locality to the first biochron of the Clarkforkian (Cf1). Archibald *et al.* (1987) estimated the range of the *Titanoides primaevus* type locality as from middle to late Tiffanian (Ti3? to Ti5?). On the basis of current studies, the Witter Locality is considered to be Ti3 or Ti4. Archibald *et al.* (1987) similarly assigned "McKenna's *Titanoides*" (Cross Locality) and Williams County *T. primaevus* (White's River Basin Survey Site 3) to Ti3?-Ti5? These localities are herein considered to be Ti3 to Ti4 in age. The assignment by Archibald *et al.* (1987) of the Riverdale Locality to Ti4?, and its unquestioned assignment by Sloan (1987) to this biochron, should be considered, at best, tenuous. The Ti3 age of the Judson and Brisbane Localities is well documented (Holtzman, 1978; Archibald *et al.*, 1987; Sloan, 1987). The only additional North Dakota mammal locality discussed by Archibald *et al.* (1987) and Sloan (1987) is the Wannagan Creek Quarry from a horizon 66 ft (20 m) below the Bullion Creek-Sentinel Butte formational contact (Erickson, 1982). Both assign the locality to the Ti4 biochron, but there is no published paleontologic evidence to support this assignment, and its stratigraphic placement suggests more likely a Ti3 age.

Although known from younger localities elsewhere (Ti5 of Wyoming), there is no North Dakota record of *Titanoides* that is younger than Ti4. *Titanoides primaevus* may occur at localities stratigraphically below White's River Basin Survey Site 3, but the material collected cannot be unquestionably assigned to this species. *Titanoides primaevus* thus appears to be middle to late Tiffanian (Ti3-Ti5).

NOTES AND ACKNOWEDGEMENTS

The collections referred to in this report are housed at the following institutions: 1) Witter Locality (United States National Museum of Natural History, original type specimen), Yale University (Princeton collection); 2) River Basin Survey Site 3 (United States National Museum of Natural History); 3) Cross Locality (American Museum of Natural History); 4) Riverdale (Field Museum of Natural History); 5 and 6) Judson and Brisbane Localities (Science Museum of Minnesota); and 7) Mott Dump Site (North Dakota Geological Survey). All quadrangle maps referred to in this paper are U.S. Geological Survey 7.5- minute series topographic quadrangles at a scale of 1:24,000, and all are in English units. Thus map and stratigraphic distances are given in feet (ft) along with their metric equivalent (m). Locality names are also referenced by a Hartman locality numbering system (e.g., L5383). The abbreviation SMM refers to the vertebrate paleontology collections of the Science Museum of Minnesota.

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THE THERMAL STATE OF THE WILLISTON BASIN IN CANADA

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ABSTRACT

Results from the analysis of over 8400 bottom-hole temperature values from petroleum exploration wells in the Williston Basin are reviewed. Estimates of geothermal gradients, thermal conductivities and heat flow values from the Mesozoic - Cenozoic and Upper Paleozoic units indicate that regional heat flow variations with depth occur and these are related to the hydrodynamics. Anomalously high geothermal gradients and heat flow values occur in the sedimentary interval below the Paleozoic erosional surface in the Weyburn area, but high values are not evident in the Mesozoic - Cenozoic sediments above. This anomaly cannot be completely explained by the blanketing effect of thick shaly units present there, and so there may be an anomalous contribution from the basement. The geothermal anomaly coincides with an electrical conductivity anomaly evident from magneto-telluric investigations. Two-dimensional numerical models with anomalous heat generation or anomalous thermal conductivity based on the structure indicated from magneto-tellurics give heat flow values that correspond to those observed. It is probable that enhanced radioactive heat generation related to mineralization occurs in the mid-upper crust resulting in the electrical and thermal anomalies.

INTRODUCTION

The Williston Basin is an intracratonic basin within the Western Canada Sedimentary Basin that is circular in shape and with over 4 km of sedimentary rocks that range in age from Cambrian to Tertiary. The Phanerozoic thickness is shown in the map of figure 1, and it is seen that the basin extends into northern United States south of Saskatchewan. In Canada, the base of the sedimentary section reaches a depth of about 3.5 km.

As part of a large project to investigate the thermal nature of the Western Canada Basin carried out over a number of years by Jones and his co-workers (see for example Lam and Jones, 1984; Jones *et al.*, 1985a; Jones *et al.* 1985b), over 8400 bottom-hole temperature values (BHT) from about 7800 petroleum exploration wells in the provinces of Saskatchewan and Manitoba have been used to examine the thermal state of the Williston Basin.

Groundwater flow has been found to significantly influence the thermal regimes of sedimentary basins and in some cases to substantially influence the migration and accumulation of hydrocarbons. The effect of hydrodynamics on the occurrence of hydrocarbons has been investigated for southern Alberta by Jones *et al.* (1986). Water motion strongly affects the thermal regime in the Williston Basin in Canada, and there is evidence that it plays a role in hydrocarbon accumulation.

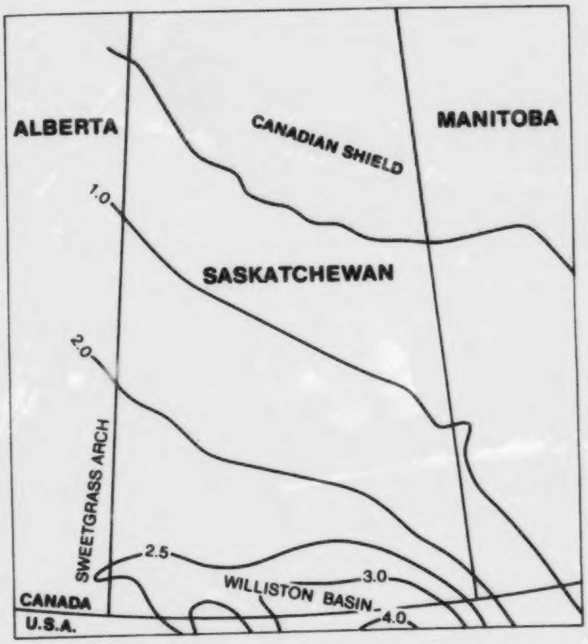


Figure 1: Phanerozoic isopach map of part of the Western Canadian Sedimentary Basin showing the Canadian part of the Williston Basin. (From Majorowicz *et al.*, 1986).

GEOTHERMAL GRADIENTS

The locations of wells from which BHT data were available are shown in Jones *et al.* (1982) and Majorowicz *et al.* (1986). The data were analyzed by Majorowicz *et al.* (1986) by grouping them into 3 x 3 township-range areas (a township-range area of the Dominion Land Survey System is approximately 9.6 km x 9.6 km). These data were plotted against depth, and the gradients were determined for two depth intervals, the Mesozoic - Cenozoic formations and the Paleozoic formations. Two intervals were chosen because major changes in conductivity exist between these two parts of the sedimentary column. The gradients in the upper interval were calculated from a least squares fitted straight line through the BHT in that interval with a fixed ground surface temperature of 5°C (Judge, 1971). The gradients below the Paleozoic surface were calculated from BHT values only by least-squares fitting of straight lines.

Since BHT values at successive times were available for some wells in the area, the correction described by Fertl and Wichmann (1977) was applied to account for the disturbance of the wells due to drilling and so determine the equilibrium formation temperatures for these wells. By comparison of the corrected and uncorrected values, a regional temperature correction of 6 percent was determined for the area and applied to all the data (see Majorowicz *et al.*, 1986).

The calculated thermal gradients are shown in figure 2 for both the upper and lower parts of the sedimentary strata.

THERMAL CONDUCTIVITIES

Effective thermal conductivities for the two depth intervals were calculated by Majorowicz *et al.* (1986) using net rock analyses and assumed heat conductivity values for the different rock types. The method was that used and described by Jones *et al.* (1984). The obtained values for the two intervals are shown in figure 3. The mean heat conductivity determined for the Mesozoic-Cenozoic formations was 2.1 $Wm^{-1}K^{-1}$ and for the Paleozoic formations was 3.7 $Wm^{-1}K^{-1}$ (Majorowicz *et al.*, 1986).

HEAT FLOW

The vertical heat flow can be determined from the product of the thermal conductivity and the vertical temperature gradient. If the flow of heat is not disturbed by water motion or other effects, the calculated heat flow will be the same in all parts of the sedimentary column. The heat flow (Q) was calculated from the estimated gradients and effective thermal conductivities for the Williston Basin in Canada for the two depth intervals (Mesozoic - Cenozoic and Paleozoic) by Majorowicz *et al.* (1986) and the results are given in Figure 4.

From figure 4 it is evident that the heat flow values calculated for the intervals above and below the Paleozoic erosional surface differ both locally and regionally. Furthermore, it is found that the mean heat flow within the Paleozoic formations is greater than that through the Mesozoic - Cenozoic formations. This, together with the evidence from the contour maps of Figures 2 and 4 that both the thermal gradient and heat flow are higher toward the northeast in the upper part of the section and the knowledge that the regional groundwater flow is northeastward (Hitchon, 1969), indicate that heat is transported by water motion northeastward across the region.

Another feature of the heat flow pattern evident in figure 4 is the anomaly in the Weyburn area in the deeper sediments but which does not occur in the upper part of the interval.

Heat flow differences between the Mesozoic-Cenozoic sediments and the Paleozoic sediments have been calculated for this area together with the rest of the Western Canada Sedimentary Basin and the result is shown in figure 5. The ΔQ ($\Delta Q_1 = Q_1 - Q_2$, where Q_1 = heat flow through the Mesozoic - Cenozoic formations and Q_2 = heat flow through the Paleozoic

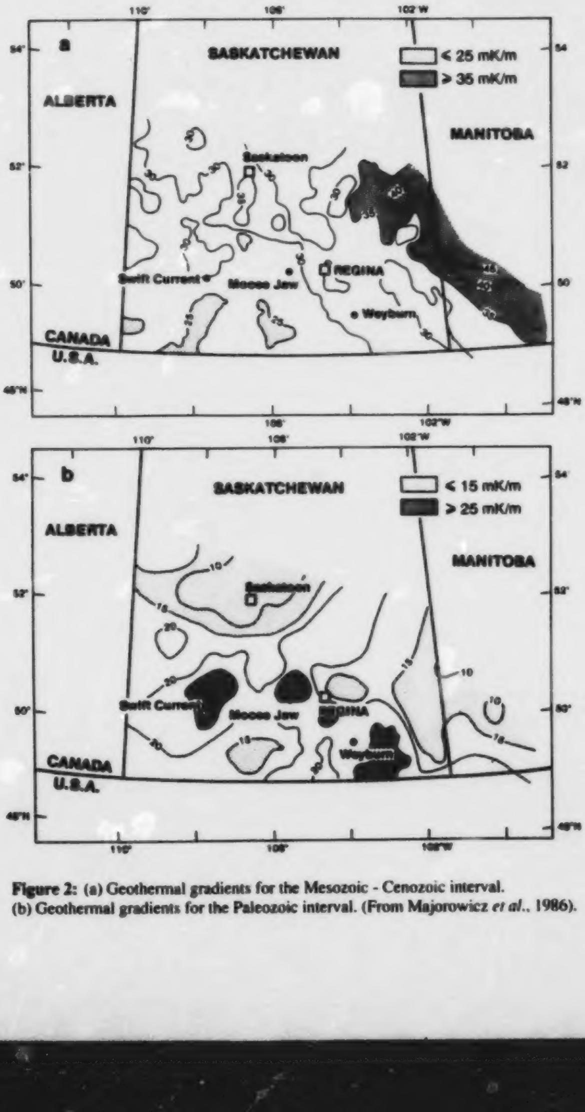


Figure 2: (a) Geothermal gradients for the Mesozoic - Cenozoic interval.
(b) Geothermal gradients for the Paleozoic interval. (From Majorowicz *et al.*, 1986).

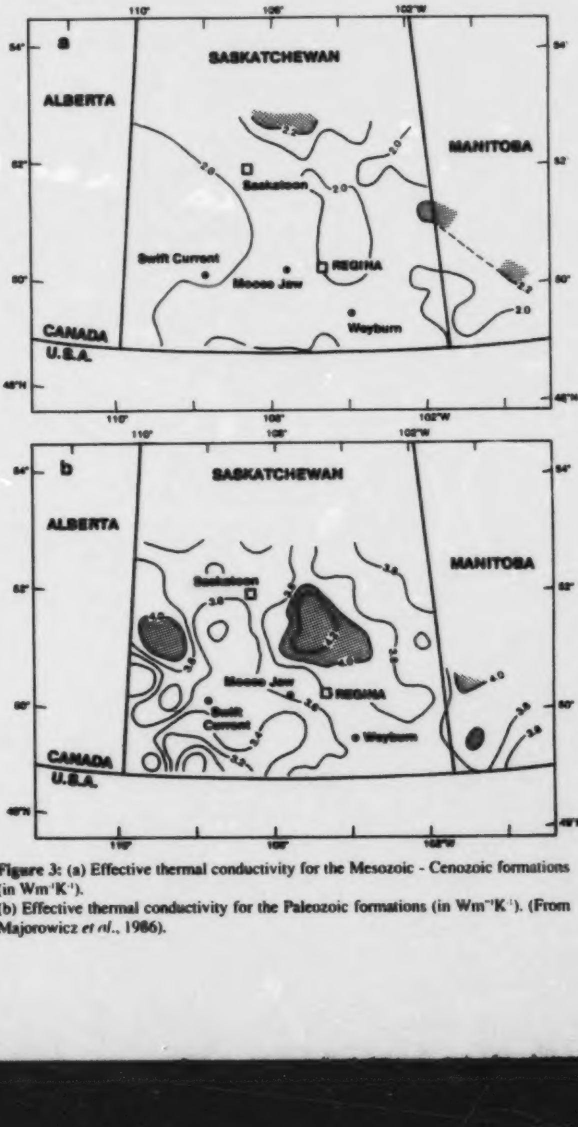


Figure 3: (a) Effective thermal conductivity for the Mesozoic - Cenozoic formations (in $Wm^{-1}K^{-1}$).
(b) Effective thermal conductivity for the Paleozoic formations (in $Wm^{-1}K^{-1}$). (From Majorowicz *et al.*, 1986).

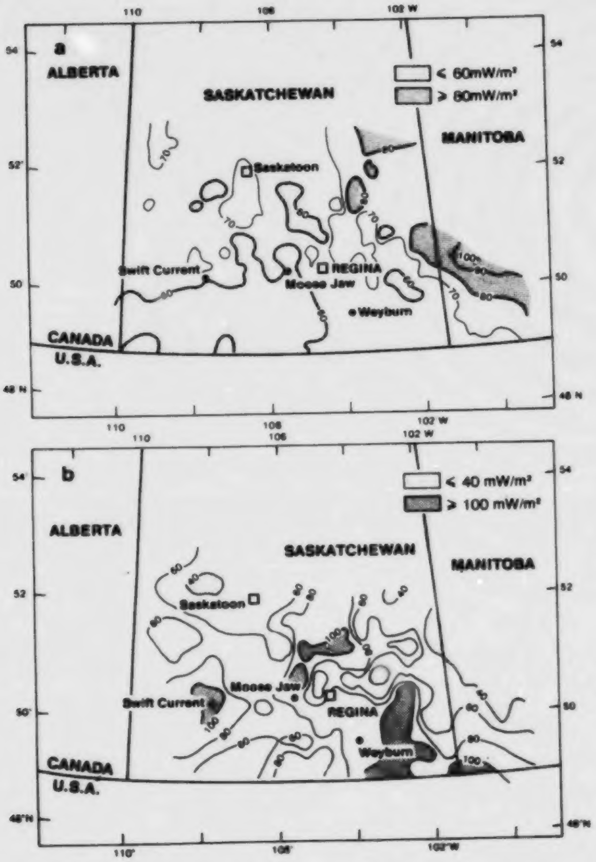


Figure 4: (a) Heat flow through the Mesozoic - Cenozoic section.
(b) Heat flow through the Paleozoic section. (Adapted from Majorowicz *et al.*, 1986).

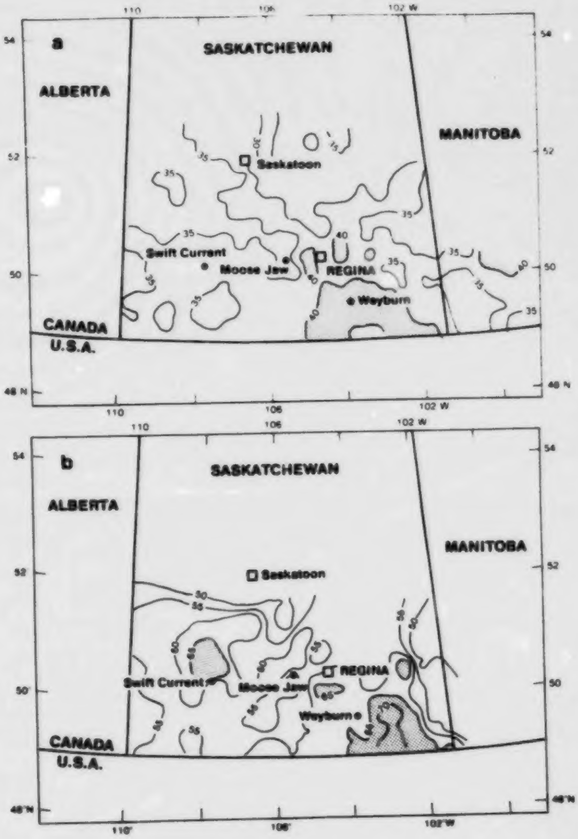


Figure 6: (a) Temperatures at 1 km depth (°C).
(b) Temperatures at 2 km depth (°C) (From Majorowicz *et al.*, 1986).

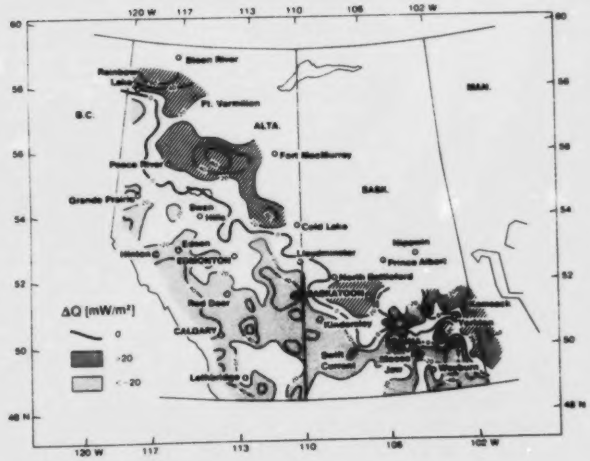


Figure 5: Contour plot of the differences $\Delta Q_i = Q_{1i} - Q_{2i}$ between the heat flow density values for the Mesozoic - Cenozoic (Q_1) and Paleozoic (Q_2) formations. (From Majorowicz *et al.*, 1986).

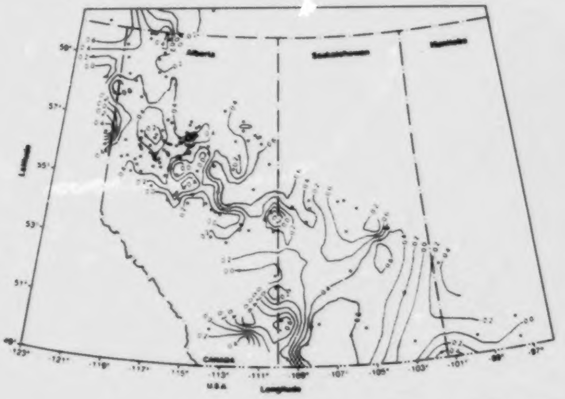


Figure 7: Contour map of log A (A = heat generation in μWm^{-3} for Precambrian basement rocks in the Western Canada Sedimentary Basin. Relatively high heat generation zones are shaded. The dots indicate locations of wells from which samples have been measured. (From Jones and Majorowicz, 1987).

formations, and the subscript "i" indicates location) pattern exhibits mainly positive values in the northeastern region and negative values to the south and southwest, with a zone where $\Delta Q = 0$ between. This is interpreted as the effect of the transport of heat by regional groundwater flow from recharge areas in the south and west upward through the porous formations of the sediments toward the discharge areas in the northeast.

THE WEYBURN ANOMALY

An area of particular interest is that around Weyburn. The temperatures at 1 and 2 km depths are shown in figure 6. High temperatures are evident in the southeast part of the area around Weyburn, and they reach more than 90°C at the basement surface there (Majorowicz *et al.*, 1986).

A southward increase in the thickness of the low conductivity shaly clastic sediments leads to a blanketing effect (see Rybach, 1981), but this does not totally account for the thermal anomaly (Majorowicz *et al.*, 1986). The most likely cause for the Weyburn anomaly is the eastward transport of heat from deep shale sediments to the west by water motion. There may be a contribution to the Weyburn high from a high temperature region in the crystalline crust. However, the heat generation map for Precambrian basement rocks in the Prairies does not indicate such (see Fig. 7, taken from Jones and Majorowicz, 1987), though it should be noted that the measured values used to construct the map of figure 7 were taken from near the top of the Precambrian basement.

Figure 8 shows the locations of known Mississippian oil pools in the Canadian part of the Williston Basin and their relation to isotherms at different depths in the Weyburn area. Water movement is the principal agent for mass and heat transfer in sedimentary basins, so correlations such as is apparent here between the temperature field and hydrocarbon occurrences are related to the hydrodynamics.

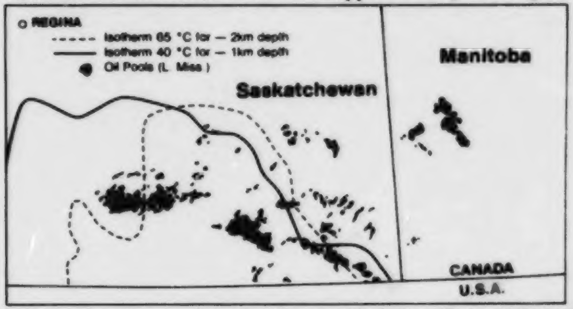
MAGNETO-TELLURIC RESULTS FROM SOUTHEASTERN SASKATCHEWAN

The electrical structure of the Williston Basin has undergone considerable study, beginning with early geomagnetic deep sounding (GDS) measurements that included the discovery and subsequent examination of the North American Central Plains (NACP) conductive structure (see Jones and Craven, 1990, for further details and the relevant references). Subsequent magneto-telluric (MT) studies were carried out by Jones and Savage (1986), Rankin and Pascal (1990) and Jones and Craven (1990).

The locations of MT profiles by Rankin and Pascal (1990) and Jones and Savage (1986) are shown in figure 9 with respect to the Weyburn heat flow anomaly. Also shown in this figure is the location of the NACP anomaly as originally placed by Alabi *et al.* (1975) as well as the apparent extension of the Weyburn heat flow anomaly into the United States as defined by Price *et al.* (1986) from S_2 pyrolysis peak studies. The work of Jones and Savage (1986) indicates that the NACP anomaly in southern Saskatchewan is at 103° W, rather than where originally indicated by Alabi *et al.* (1975), and so it coincides with the Weyburn heat flow anomaly.

Jones and Craven (1990) examined the correlation of the NACP electrical conductivity anomaly with other geophysical data. They found that the MT data along MT2 (see Fig. 9), when modelled in 2D, are consistent with a high conductivity single structure or series of structures that lie in a relatively thin (1-5 km thick) arc with the top at a depth of 10 km and centered on 103° W. Two models that they deduce from their work are given in figure 10.

Jones and Craven (1990) investigated whether or not the gravity data along the profile MT2 could be modelled by bodies of the same general character as those imaged by the MT data. Their results are illustrated in figure 11 taken from their paper. The arcurate shape of the NACP anomaly suggested by the MT results and adapted to the gravity model can account for the observed gravity anomaly at 103° W. It is interesting to note though, that a broad gravity anomaly exists to the west of the "revised" location of the NACP - at approximately the original location of the NACP anomaly as indicated by Alabi *et al.* (1975). Is it possible that the original GDS data of Alabi *et al.* (1975) detected a deeper conductive structure but the NACP anomaly as now recognized by Jones and Savage (1986) and Jones and Craven (1990) masked its presence? Have the MT data of Jones and Craven not detected a deep structure to the west of the NACP anomaly?



o REGINA
Isotherm 65 °C for — 2km depth
Isotherm 40 °C for — 1km depth
Oil Pools (L. Miss.)
Saskatchewan
Manitoba
CANADA
U.S.A.

Figure 8: Oil pool locations in the Canadian part of the Williston Basin and observed temperatures. (From Majorowicz *et al.*, 1986).

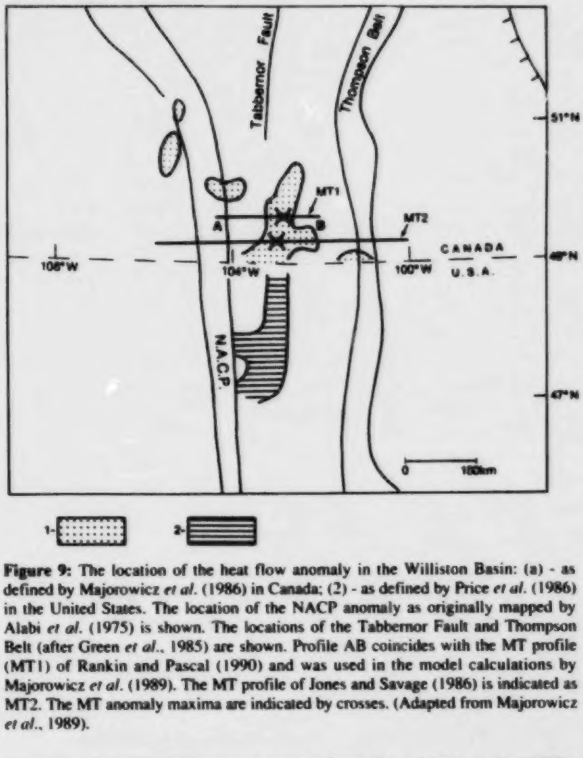


Figure 9: The location of the heat flow anomaly in the Williston Basin: (a) - as defined by Majorowicz *et al.* (1986) in Canada; (2) - as defined by Price *et al.* (1986) in the United States. The location of the NACP anomaly as originally mapped by Alabi *et al.* (1975) is shown. The locations of the Tabbernor Fault and Thompson Belt (after Green *et al.*, 1985) are shown. Profile AB coincides with the MT profile (MT1) of Rankin and Pascal (1990) and was used in the model calculations by Majorowicz *et al.* (1989). The MT profile of Jones and Savage (1986) is indicated as MT2. The MT anomaly maxima are indicated by crosses. (Adapted from Majorowicz *et al.*, 1989).

HEAT FLOW MODELLING

Majorowicz *et al.* (1989) used the two-dimensional numerical finite-difference method of Jones and Ertman (1985) to investigate the possible character of the thermal region associated with an anomalous mid-upper crustal structure derived from the MT results from Jones and Savage (1986) and Rankin and Pascal

(1990). An anticlinal structure was modelled, and two possible causes for the thermal anomaly were considered - either anomalous heat generation or anomalous thermal conductivity. Because it was not possible to fit a structure with uniform properties to the observed heat flow data, the structure was divided into sections of different anomalous heat generation values DA or different thermal conductivity values K.

Good fits to the observed values were obtained for two models as shown in figures 12a, b. The heat generation model (Fig. 2a) exhibits anomalous heat generation in the mid-upper crust for three of the structural sections of 2.2, 4.0 and 7.5 μWm^{-3}, with the largest section having value 4.0 μWm^{-3}. Heat generation in Precambrian basement rocks can vary between 0.1 μWm^{-3} to 14 μWm^{3} (Beach *et al.*, 1987 and Jones and Majorowicz, 1987), and in most heat-generation models the heat generation decreases with depth. However, the situation in the Williston Basin may reflect widespread redistribution of radiogenic material from oceanic sediments overriding Superior crust in the Superior - Churchill boundary zone (Drury, 1985).

A thermal conductivity model that fits the observations is shown in figure 12b. The thermal conductivity of the structure is more than three times that of the overlying strata. Also, it is interesting to note that at approximately 70 - 75 km west of the peak of the anticlinal structure at the location of the NACP anomaly as originally placed by Alabi *et al.* 1975), there is a shoulder on the main structure.

It appears that the more probable cause for the anomalous heat flow is enhanced heat generation in the mid-upper crust related to increased abundances of radiogenic uranium, thorium and potassium as suggested by the model of figure 12a (Majorowicz *et al.*, 1989). Such enhancement of radiogenic elements could accompany high electrical conductivity mineralization in the anomalous zone.

It should be noted here that a number of differences exist between the observed results from the BHT data and the thermal model. The heat flow anomaly in the Weyburn area is apparent only within the Paleozoic sediments (see Fig. 4). However, the model is constructed as if the surface of the Earth is at the Paleozoic surface (and maintained at 0°C). Furthermore, although the initial temperature distribution was taken as 25°C km^{-1}, the temperature versus depth distributions after 15 Ma (the time of the development of the thermal fields) at the sides of the model are different from this. To the east, the average temperature gradient is about 19°C km^{-1} which, from figure 4, appears to be nearly consistent with the observed data. However, to the west, the model results indicate a final average gradient of more than 30°C km^{-1}. This latter value is of the order of 30-40 percent greater than the apparent value indicated in figure 4. The heat flow estimates have an uncertainty of perhaps 25 percent (see Majorowicz *et al.*, 1986). The reason for this inconsistency is not clear. It appears that for various reasons the area to the west of the main anomaly should be considered further.

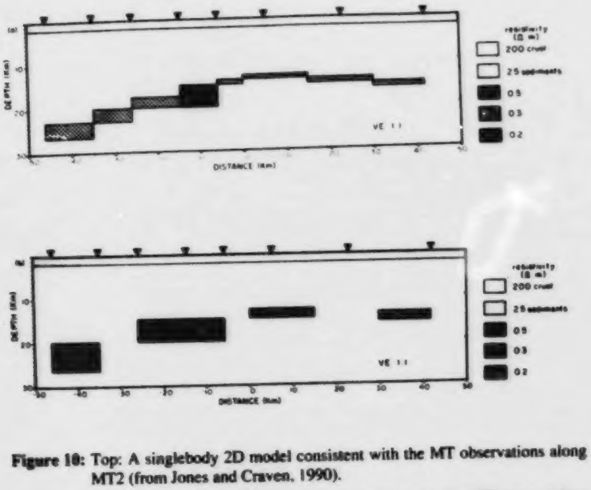


Figure 10: Top: A singlebody 2D model consistent with the MT observations along MT2 (from Jones and Craven, 1990).

Bottom: A multibody 2D model consistent with the MT observations along MT2 (from Jones and Craven, 1990).

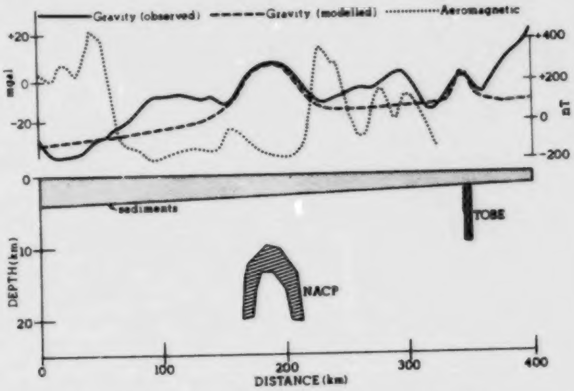


Figure 11: Upper part: Gravity (solid line) and aeromagnetic (dotted lines) observed data along profile MT2 compared with the response (dashed line) to the gravity model of the lower part of the figure.

Lower part:A model that explains two of the observed anomalies in the gravity data. The sediments are modelled with a density contrast of -180 kg m^{-3} the NACP with +270 kg m^{-3}, and the TOBE with +180 kg m^{-3}. (From Jones and Craven, 1990).

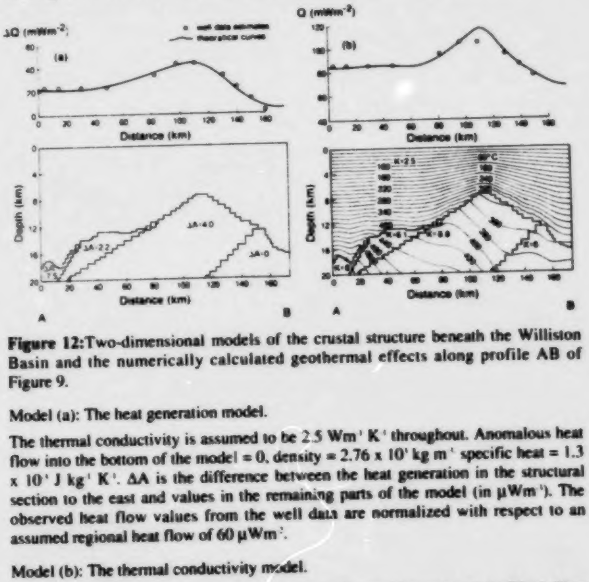


Figure 12: Two-dimensional models of the crustal structure beneath the Williston Basin and the numerically calculated geothermal effects along profile AB of Figure 9.

Model (a): The heat generation model.

The thermal conductivity is assumed to be 2.5 Wm^{-1} K^{-1} throughout. Anomalous heat flow into the bottom of the model = 0, density = 2.76 x 10^{3} kg m^{-3} specific heat = 1.3 x 10^{3} J kg^{-1} K^{-1}. ΔA is the difference between the heat generation in the structural section to the east and values in the remaining parts of the model (in μWm^{-3}). The observed heat flow values from the well data are normalized with respect to an assumed regional heat flow of 60 μWm^{-2}.

Model (b): The thermal conductivity model.

The heat generation is assumed uniform throughout and equal to 1.4 μWm^{-3}. Heat flow into the bottom of the model is 20 μWm^{-2} and the density and specific heat are as in model a. Assumed thermal conductivities, K (in Wm^{-1} K^{-1}), are assigned to each structural section. Isotherms are in °C. The upper boundary in both models was maintained at 0°C and a temperature variation of 25°C km^{-1} was used as the initial starting condition. The temperature gradients across the side boundaries were zero, so no heat flux was permitted to flow laterally at these boundaries. A time of 15 million years in 30 x 10^{3} year increments was allowed for the thermal field to develop.

CONCLUSIONS

The bottom-hole temperature data from the Williston Basin indicate that regional heat flow variations with depth occur and these are likely related to the hydrodynamics and it appears that hydrocarbon accumulations are related to the hydrodynamics and hence the thermal regime. A temperature and heat flow anomaly occurs in the Weyburn area, and is coincident with a magnetotelluric electrical conductivity anomaly. The evidence suggests that enhanced radioactive heat generation related to mineralization occurs in the mid-upper crust, and this produces both the thermal and electrical anomalies.

ACKNOWLEDGEMENTS

The work in compiling this review has been supported by the Natural Sciences and Engineering Research Council of Canada.

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STRUCTURAL DEVELOPMENT OF THE WILLISTON BASIN IN SOUTHWESTERN NORTH DAKOTA

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ABSTRACT

A total of 40 formations were picked from approximately 2200 oil well wireline logs. Isopach maps and structure contour maps were constructed from these data for each of the formations. Individual structural features were delineated through a combination of trend surface and residual analysis. Analysis of the isopach maps and of structural relief permitted the determination of the timing of tectonic events for the individual structures and the area as a whole.

In addition to the major structures, several less prominent structures, both positive and negative, have been identified. Although the subsidence of the basin was the principal regional tectonic activity, the structural features in southwestern North Dakota were intermittently decoupled from the larger-scale basin movement, producing local highs and lows. All of these structural features appear to have behaved independently of each other through time. Basin initiation and subsequent subsidence are the only recognizable pre-Devonian activity in the area. Most of the development of the individual structures took place during the Devonian and Mississippian, although the relief on some structures was enhanced by Laramide activity at the end of the Cretaceous.

INTRODUCTION

The Williston Basin in southwestern North Dakota contains several prominent structural features, including the Cedar Creek, Billings, and Little Knife anticlines, as well as the southern end of the Nesson anticline. The purpose of this paper is to document the development of the basin and its individual structural features over much of the southwestern part of the state.

PREVIOUS WORK

Numerous large-scale studies of the structure and tectonics of the Williston Basin region have been completed (Carlson, 1960; Anderson and Hunt, 1964; Macauley, 1964; Carlson and Anderson, 1965; Parker, 1967; Thomas, 1974; Brown, 1978; Sheldon and Carter, 1979; Caldwell, 1982; Gerhard *et al.*, 1982; Shurr, 1982; Peterson, 1984; Peterson *et al.*, 1985; Maughan and Perry, 1986; Brown and Brown, 1987; Peterson and MacCary, 1987; Shurr *et al.*, 1989), but little work has been done which relates specifically to the development of individual structures within the area of study. The majority of published papers dealt primarily with oil field characteristics as they relate to production and development (*e.g*, Martens, 1978; Wittstrom and Hagemeier, 1978; Kupecz, 1984; Narr and Burrus, 1984; Bogle and Hanson, 1987; DeMis, 1987; Lindsay, 1987; Sturm, 1987; Breig, 1988; Burke, 1990).

METHODS

The area of study extends from T133N to T149N, and from longitude 102W to the North Dakota-Montana state line (Fig. 1). The area thus is roughly square, and approximately 10,000 sq mi (25,000 km^2), and includes all of the well-known structural features in that part of the basin, with the exception of the Cedar Creek anticline.

Data were collected from a total of about 2200 wireline logs. The only wells in the study area which were not included were directional wells or those in fields with a high density of wells; in the latter case, logs were examined only if the well was more than 0.5 miles (0.8 km) from the closest well control. Altogether, 40 formation tops were picked from the logs (Tables 1 and 2). The tops were used to construct structure contour and isopach maps. Control points for the top of the Charles Formation are depicted in Figure 2.

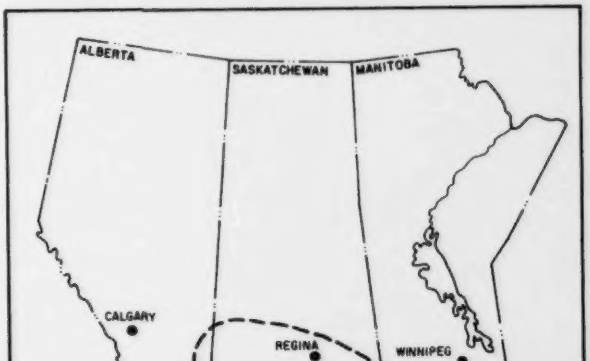

Figure 1: Location map of study area (shaded) within Williston Basin.

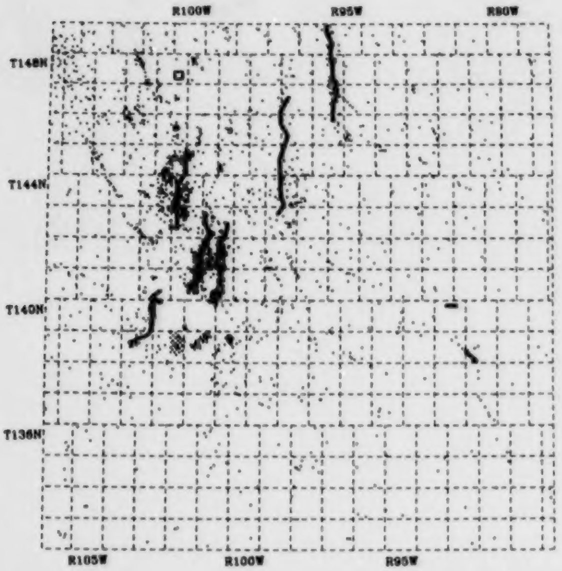

Figure 2: Distribution of control points for the top of the Charles Formation. Dashed lines are township boundaries; heavy lines indicate position of structural features described in the text. Townships are 6 mi (9.6 km) across.

Initial identification of individual structural elements was accomplished through polynomial trend-surface analysis. For each conformable top, a trend surface was calculated, which yielded the regional basin surface. Subtraction of the actual data from the calculated surface was done to obtain a map of the deviations (residuals). Positive residuals indicate areas which stand higher than the regional trend and may be positive structural features; the reverse is true of the negative residuals. Areas identified by residual analysis were then examined on the structure contour and isopach maps to determine whether or not they showed any evidence of structural activity, and to refine the definition of the structural areas.

Table 1: Stratigraphic tops used in this study and the number of wells penetrating them.

Formation	No. of Wells	Formation	No. of Wells
Pierre	508	Charles	1864
Niobrara	1027	Lodgepole	1054
Greenhorn	2157	Bakken	868
Mowry	2164	Three Forks	1000
Inyan Kara	2113	Bird Bear	944
Swift	2052	Duperow	946
Rierdon	1809	Souris River	826
Spearfish		Dawson Bay	668
Saude Mbr	1722	Prairie	662
Pine Salt Mbr	1593	Winnipegosis	650
Belfield Mbr		1745Ashern	643
Minnekahta	2059	Interlake	666
Opeche	2065	Stonewall	653
Minnelusa	2100	Stony Mountain	654
Tyler	2099	Red River	654
Otter	1927	Winnipeg Gp.	
Kibbey		Roughlock	94
sandstone mbr	1728	Icebox	96
limestone mbr	1729	Black Island	94
shale mbr	1728	Deadwood	76
		Precambrian 12	

Table 2: Isopach maps used in this study and the number of wells from which they are constructed.

Formation	No. of Wells	Formation	No. of Wells
Pierre	465	Madison Gp.	994
Niobrara-Carlile	1027	Charles-Mission	
Greenhorn-Belle		Canyon	1039
Fourche	2147	Lodgepole	995
Mowry-Newcastle	2102	Bakken	862
Inyan Kara	2045	Three Forks	933
Swift	1792	Birdbear	936
Rierdon-Piper	1720	Duperow	819
Spearfish	1744	Souris River	665
Saude Mbr	1610	Dawson Bay	666
Pine Salt Mbr	1591	Prairie	659
Belfield Mbr	1743	Winnipegosis	649
Minnekahta	2055	Ashern	642
Opeche	2060	Interlake	652
Minnelusa	2091	Stonewall	653
Tyler	1926	Stony Mountain	653
Otter	1847	Red River	96
Kibbey	1851	Winnipeg Gp.	76
sandstone mbr	1724	Roughlock	94
limestone mbr	1727	Icebox	95
shale mbr	1725	Black Island	75
		Deadwood	12

Once tectonically active areas were identified, the history of movement for each area was examined with structural relief plots. The plots illustrate the present difference in elevation between points on the structure and nearby points off the structure at a given horizon. The relief reflects the difference in vertical motion between the on- and off-structure locations since deposition of bed. Thus, continued relative uplift of a structure should result in a relief curve which rises with increasing age, and continued subsidence should produce the opposite effect. Factors other than tectonic activity, such as local paleotopography or differential compaction, may also affect the structural relief. Therefore, the relief diagrams and the numbers taken from them should be regarded as correct in showing direction and magnitude of tectonic movement, but not as precise indicators of tectonic uplift or subsidence, especially on a small scale. In the descriptions below, unless otherwise stated, uplift and subsidence refer to the relative movement of the presently high-standing block.

In the calculation of tectonic subsidence, we used backstripping techniques (*e.g.*, Sclater and Christie, 1980) to calculate the tectonic component of subsidence by decompaction of sedimentary units to their thicknesses immediately after deposition, and by subtraction of the subsidence due to sediment loading. For sea level correction, we used the Phanerozoic sea level curves of Vail *et al.* (1977) and Haq *et al.* (1987). We have also followed the suggestion of Bond and Kominz (1984) and expressed the subsidence in terms of two models, for minimum and maximum compaction of the sediments. This yields two curves on tectonic subsidence plots, which represent the extreme cases in porosity loss and provide limits on the subsidence models.

The relative ages of the units were taken from Ballard *et al.* (1983); the corresponding absolute ages are from Harland *et al.* (1981) (Table 3).

Table 3: Estimated ages of the tops of the units used in this study, after Ballard *et al.* (1983) and Harland *et al.* (1981).

Formation	Age (Ma)	Formation	Age (Ma)
Pierre	70	Charles	341
Niobrara	82	Lodgepole	355
Greenhorn	91	Bakken	362
Mowry	100	Three Forks	370
Inyan Kara	107	Birdbear	377
Swift	154	Duperow	381
Rierdon	160	Souris River	383
Spearfish		Dawson Bay	385
Saude Mbr	246	Prairie	386
Pine Salt Mbr	250	Winnipegosis	387
Belfield Mbr	252	Ashern	389
Minnekahta	266	Interlake	409
Opeche	269	Stonewall	422
Minnelusa	275	Stony Mtn	432
Tyler	315	Red River	440
Otter	332	Winnipeg Group	
Kibbey		Roughlock	446
sandstone	mbr 335	Icebox	447
limestone mbr	337	Black Island	450
shale mbr	339	Deadwood	491

RESULTS

DELINEATION OF AREAS

By the means described above, we were able to delineate eight areas which showed significant structural activity (Fig. 3). Three features that lie within the area of study are not included in this analysis: 1) the southern end of the Nesson anticline, which has been covered in some detail in a previous work (LeFever *et al.*, 1987); 2) the structure under MonDak field, which straddles the North Dakota-Montana state line and could not be delineated adequately with the data we had; and 3) the Red Wing Creek structure, which we consider to be an impact feature, and not relevant to the structural evolution of the basin; the position of this structure is shown on all maps by an open square.

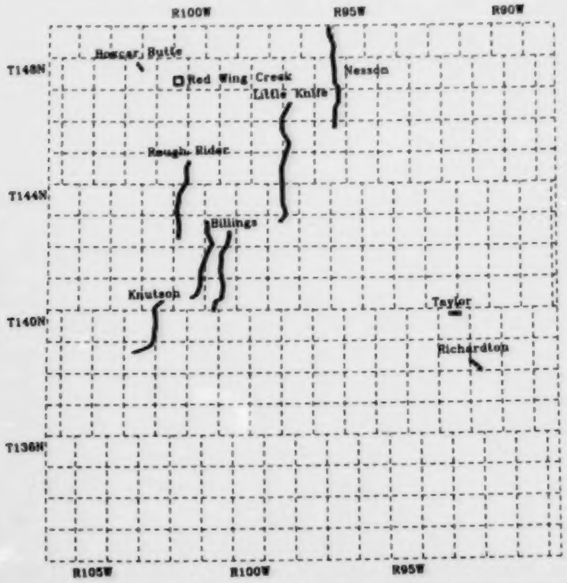

Figure 3: Locations of structural features described in the text. Dashed lines are township boundaries; townships are 6 mi (9.6 km) across.

Structure contour and isopach maps for selected units are depicted in Figures 4 and 5. The structure on the Charles Formation for the individual areas is depicted in Figure 6.

Boxcar Butte High

The Boxcar Butte structure, located in T148N, R102W, was first described by Martens (1978), largely on the basis of seismic data.

The structure is slightly elongated northwesterly and shows about 25 ft (8 m) of relief on the Mowry Formation (Cretaceous), 40 ft (12 m) on the Madison Group (Mississippian), 50 ft (15 m) on the Duperow Formation (Devonian), and about 60 ft (18 m) on the Red River Formation (Ordovician).

Despite the small amplitude, some evidence of vertical movement can be seen on the isopach maps. Thinning is shown in the isopach maps of the Greenhorn-Belle Fourche, Duperow, Dawson Bay, and Interlake Formations.

The structural relief plot (Fig. 7a) shows very little change until the Devonian. From the end of the Ashern (389 Ma) until the end of Otter (332 Ma), about 40 ft (12 m) of relief was generated. Little relative movement occurred from then until the end of the Pine Salt Member of the Spearfish Formation (250 Ma), when about 30 ft (9 m) of subsidence occurred. From then until the beginning of the Cretaceous units, about 70 ft (20 m) of uplift occurred. During the Cretaceous, about 30 ft (9 m) of renewed subsidence took place as the basin was tilted to the west.

Rough Rider Anticline

The Rough Rider anticline is located in T143N - T145N, R101W. The characteristics of fields along the structure were described by DeMis (1987) and Petty (1988); the structural history was studied in detail by Crashell (1991).

The anticline is a northward-plunging asymmetric fold, the eastern limb of which dips more steeply than the western. There is a maximum 60 ft (18 m) of relief across the northern end of the structure on the Mowry Formation, 120 ft (37 m) on the Madison Group, 150 ft (45 m) on the Duperow, and about 200 ft (60 m) on the Red River Formation. To the south, the anticline is somewhat less well-defined, and has less relief across it.

Thinning is evident along the Rough Rider anticline on many of the isopach maps (*e.g.*, Figs. 5b, 5d). The plot of structural relief (Fig. 7b) shows that significant differential movement did not begin until sometime after deposition of the Stonewall Formation (422 Ma). Uplift continued through the Devonian and the Mississippian until the end of the Lodgepole (355 Ma). At that time, the Rough Rider anticline subsided until the end of the Otter (332 Ma). The subsidence may represent reversal of movement along a bounding fault, a tilting of the area in the on-structure direction (westward), or possibly salt solution and collapse under the anticline. Uplift was renewed after the Tyler (315 Ma), and ceased prior to deposition of the Opeche Formation (269 Ma). Little vertical movement took place until after the end of Spearfish deposition (246 Ma), when about 100 ft (30 m) of uplift occurred prior to the end of Rierdon deposition (160 Ma). This uplift was followed by subsidence until sometime prior to the end of Inyan Kara deposition (107 Ma). From Inyan Kara through Niobrara time (82 Ma), about 50 ft (15 m) of uplift occurred. This movement may have been in response to Laramide activity, but, because the off-structure wells are to the east of the structure, uplift was due to activity on the anticline and not to the general westward tilt of the basin at that time.

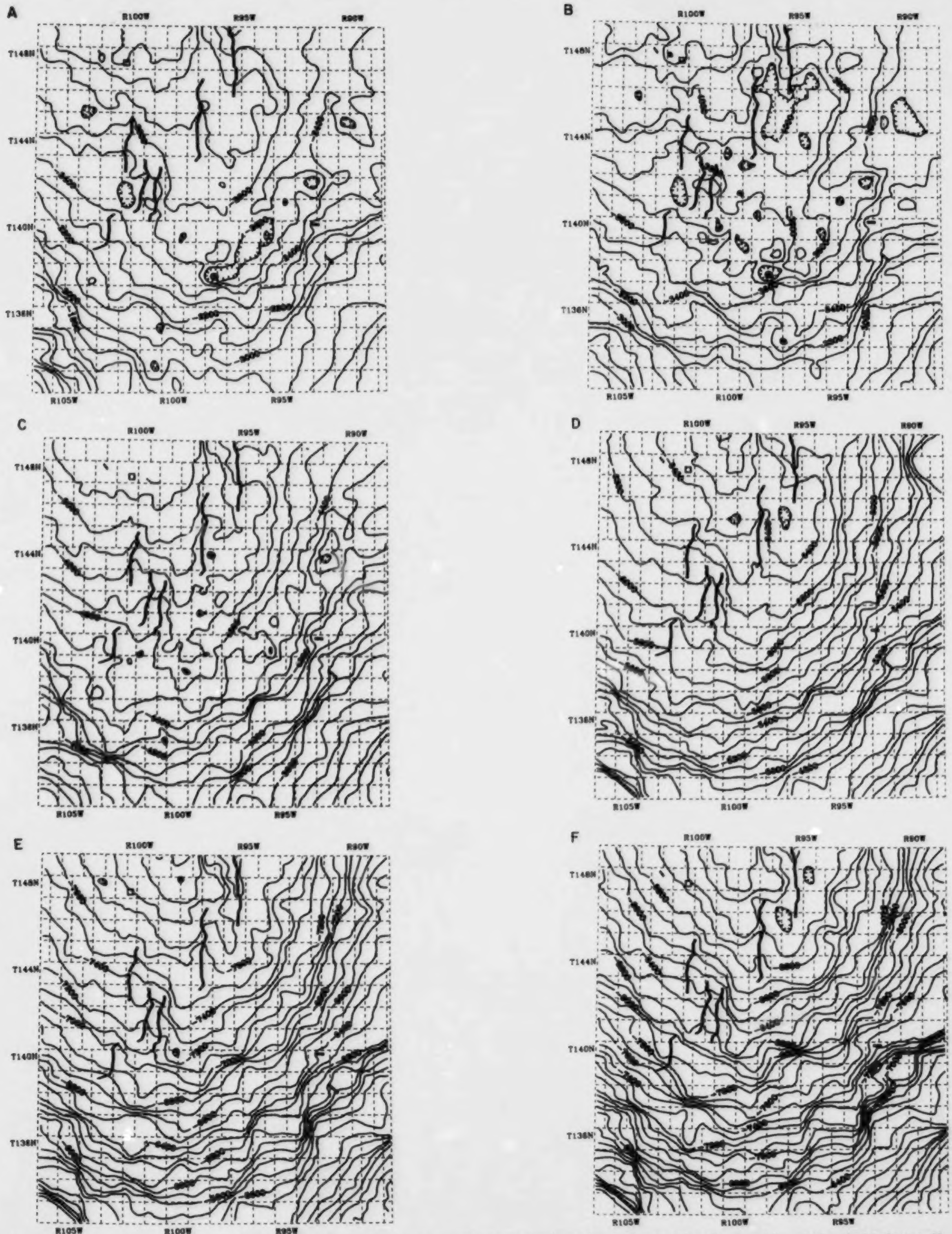

Figure 4, a) to f): Structure contour maps at selected horizons. Dashed lines are township boundaries; heavy lines indicate position of structural features described in the text. CI = 100 ft (30.5 m). Townships are 6 mi (9.6 km) across: a) Mowry Formation; b) Rierdon Formation; c) Minnekahta Formation; d) Charles Formation; e) Lodgepole Formation; f) Duperow Formation.

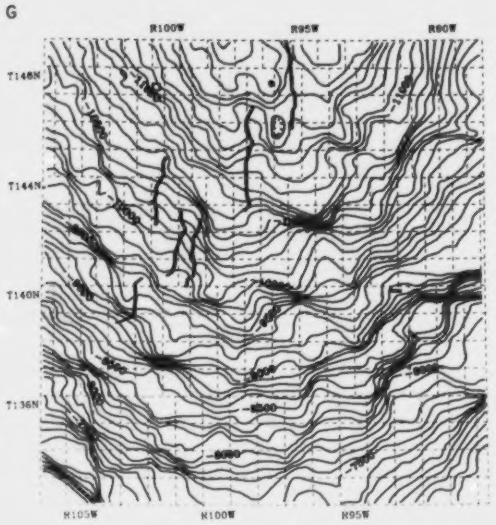

Figure 4 g): Red River Formation

Billings Anticline

The Billings anticline was described by S.B. Anderson in 1966. The fields along it have been studied by several workers (*e.g.*, Breig, 1988; Burke, 1989; Kupecz, 1984; Petty, 1988), and the structural history was studied in detail by Crashell (1991). The structure is located in T141N-T144N, and R100W-R101W. The Billings anticline is composed of two parallel folds, both of which trend slightly east of north. The southern limits of the anticline are poorly defined due to the absence of wells in the southern unit of Theodore Roosevelt National Park. The anticline may continue southward as far as T139N, into the Medora and Fryburg fields. There is a high associated with western Fryburg field that may be continuous with the eastern limb of the Billings anticline, and that of Medora field, may be continuous with the western limb.

Closely spaced contours along the R99W - R100W boundary from T140N to T144N suggest that the anticline is fault-bounded on its eastern side. The anticline shows about 80 ft (25 m) of relief on the Mowry Formation, about 150 ft (45 m) on the Madison, 250 ft (75 m) on the Duperow Formation, and about 400 ft (120 m) on the Red River Formation. Noticeable thinning is present in the Mowry, Charles-Mission Canyon, Lodgepole, Duperow, Souris River, Dawson Bay, and Interlake Formations.

The plot of structural relief (Fig. 7c) indicates that little relative movement took place during the Ordovician, but from the Ashern Formation (389 Ma) to the end of the Charles Formation (341 Ma), about 150 ft (45 m) of relief developed on the structure. Between the end of Otter (332 Ma) and the beginning of the Opeche (272 Ma), another 80 ft (25 m) of relief was added. Little relief appears to have developed from the Opeche to the end of Spearfish deposition (246 Ma). Between the end of Spearfish and the end of Rierdon deposition (160 Ma), about 80 ft (25 m) of uplift occurred. During the Cretaceous, relief on the structure declined about 40 ft (12 m) from Inyan Kara to Niobrara deposition (107 - 82 Ma).

Little Knife Anticline

The Little Knife anticline is a northward-plunging fold located in T143N - T144N, R98W, and T145N - T147N, R97W. The anticline was described by Wittstrom and Hagemeier (1978), and Little Knife field has been the subject of several studies (*e.g.*, Lindsay, 1987; Narr and Burruss, 1984). Apparently not reported in the literature, a positive feature around Lone Butte field in T147N, R97W appears to be a northward extension of the structure at Little Knife field. Because the characteristics of the two are similar, we are treating the Lone Butte high as the northern end of the Little Knife anticline.

There is about 50 ft (15 m) of relief on the Little Knife anticline at the top of the Mowry Formation, 70 ft (21 m) of relief on the Madison, 100 ft (30 m) on the Duperow, and about 120 ft (36 m) on the Red River Formation. Thinning is evident in the Greenhorn-Belle Fourche, Mowry, Inyan Kara, and Charles-Mission Canyon Formations, and perhaps in the Bakken, Three Forks, Duperow, Prairie, and Interlake Formations as well, although there is very little deep control.

The relief plot of the southern part of the anticline (Fig. 7d) shows little Ordovician relief. Beginning with deposition of the Ashern Formation (389 Ma), considerable uplift occurred, about 85 ft (26 m) in about 5 million years. The rate of uplift then slowed somewhat until the post-Lodgepole (355 Ma), when an additional 100 ft (30 m) of relief was added by the end of deposition of the shale member of the Kibbey Formation (339 Ma); from then until the end of the Otter (332 Ma), little relative movement took place. From 332 Ma until the end of the Minnekahta (266 Ma), the anticline subsided about 60 ft (18 m). From 266 Ma to the end of the Mowry (100 Ma), some 80 ft (25 m) of uplift occurred, followed by about 25 ft (8 m) of subsidence by the end of Niobrara time (82 Ma).

The northern end of Little Knife anticline shows a generally similar history but with two differences. First, the relief is less, usually about half that farther south, and the rate of change of relief is less. Second, the interval between Otter and Minnekahta deposition is one of minor uplift instead of subsidence. Thus, if these two areas are over the same structure, southward tilting is implied.

Knutson Anticline

This structure is a northward-plunging fold located under Knutson field in T140N, R103W, and extends south under Square Butte field in T139N, R103W (Bogle and Hansen, 1987).

There is about 20 ft (6 m) of relief across the Knutson anticline on the Mowry Formation, about 30 ft (9 m) on the Madison Group, 50 ft (15 m) on the Duperow Formation, and about 60 ft (18 m) on the Red River Formation. Definition of the structure is hampered by the almost complete lack of control below the Madison Group, and immediately to the east in the south unit of Theodore Roosevelt National Park. Few units appear to thin across the Knutson anticline.

The plot of structural relief (Fig. 7e) shows little activity, but control is so poor at depth that off-structure wells somewhat basinward of the structure were used. This ought to have the effect of increasing the calculated relief by the addition of the general basin subsidence. No significant relief is apparent at the end of the Stonewall Formation (422 Ma). From 389 Ma to 386 Ma (Ashern - Prairie Formations) about 10 ft (3 m) of relief developed, about 10 ft (3 m) more from 385 Ma to 370 Ma (Dawson Bay - Three Forks Formations), and 15 ft (5 m) between Three Forks and Lodgepole.

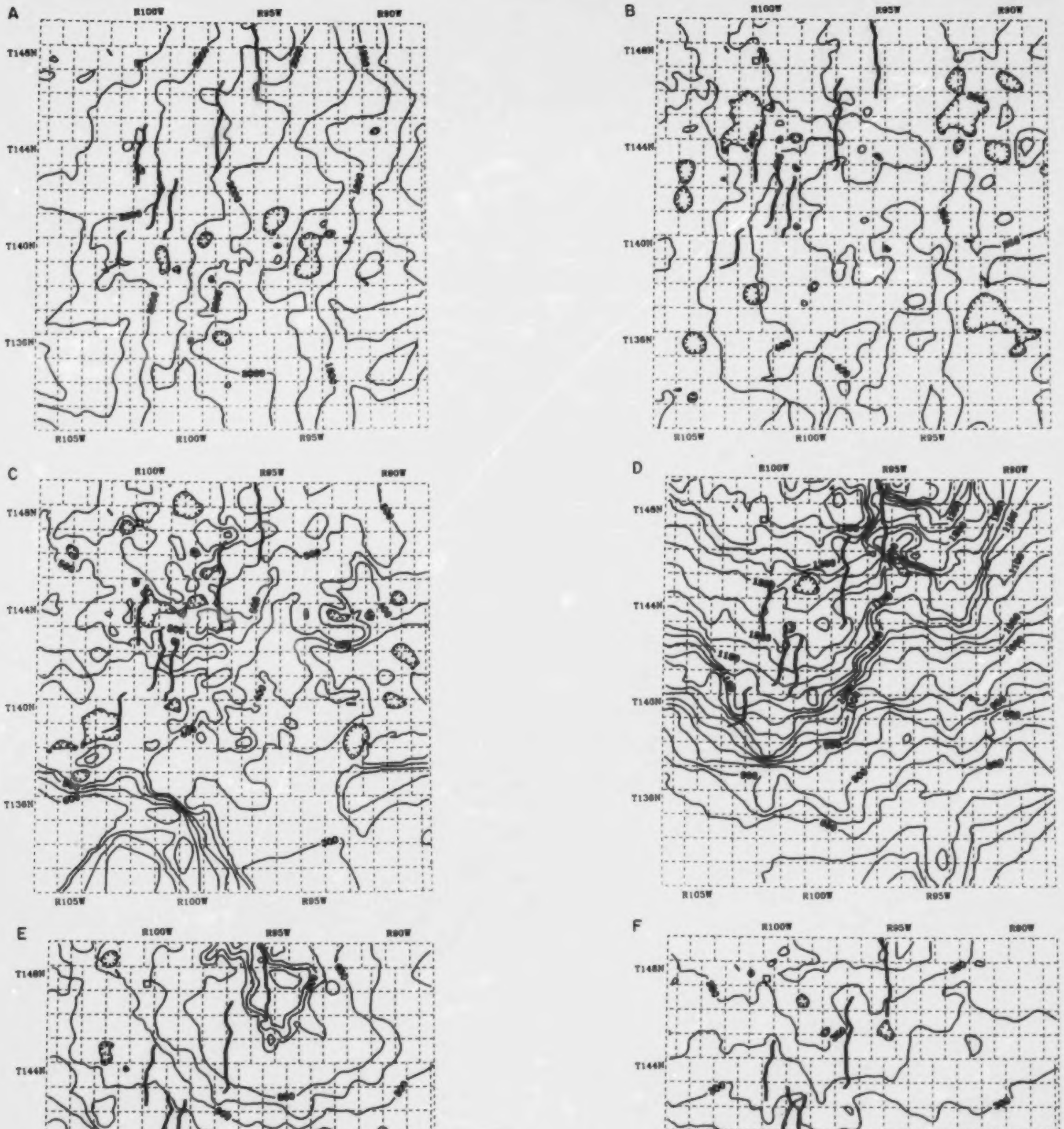

Figure 5: Isopach maps of selected formation. Dashed lines are township boundaries; heavy lines indicate position of structural features described in the text. Townships are 6 mi (9.6 km) across: a) Pierre Formation; CI = 100 ft (30.5 m); b) Mowry and Newcastle Formations; CI = 25 ft (7.6 m); c) Rierdon and Piper Formations; CI = 50 ft (15.2 m); d) Charles and Mission Canyon Formations; CI = 25 ft (7.6 m); e) Lodgepole Formation; CI = 25 ft (7.6 m); f) Duperow Formation; CI = 25 ft (7.6 m)

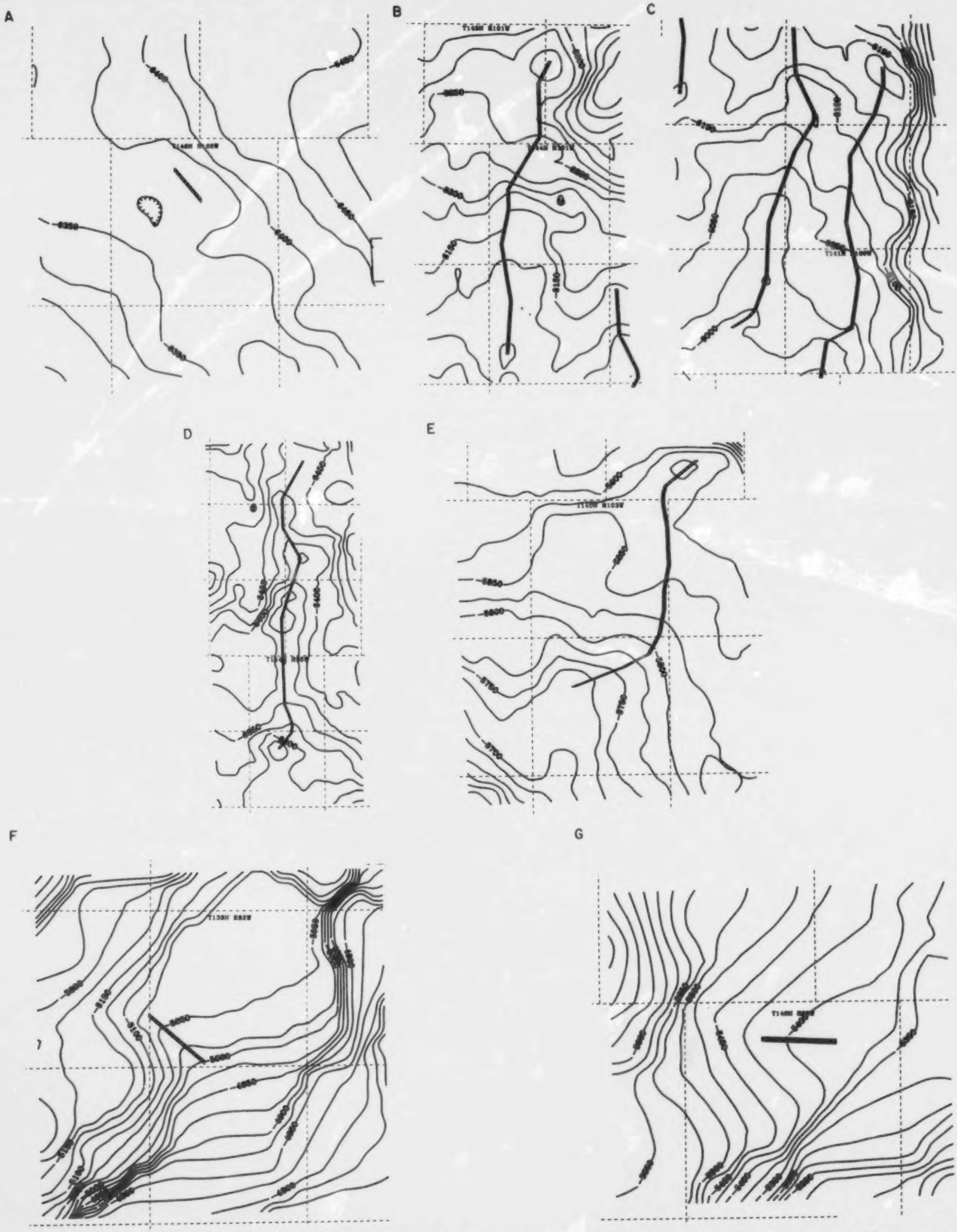

Figure 6: Structure contour maps on the Charles Formation for individual areas. Dashed lines are township boundaries. CI = 25 ft (7.6 m). Townships are 6 mi (9.6 km) across: a) Boxcar Butte High; b) Rough Rider Anticline; c) Billings Anticline; d) Little Knife Anticline; e) Knutson Anticline; f) Richardton High; g) Taylor High

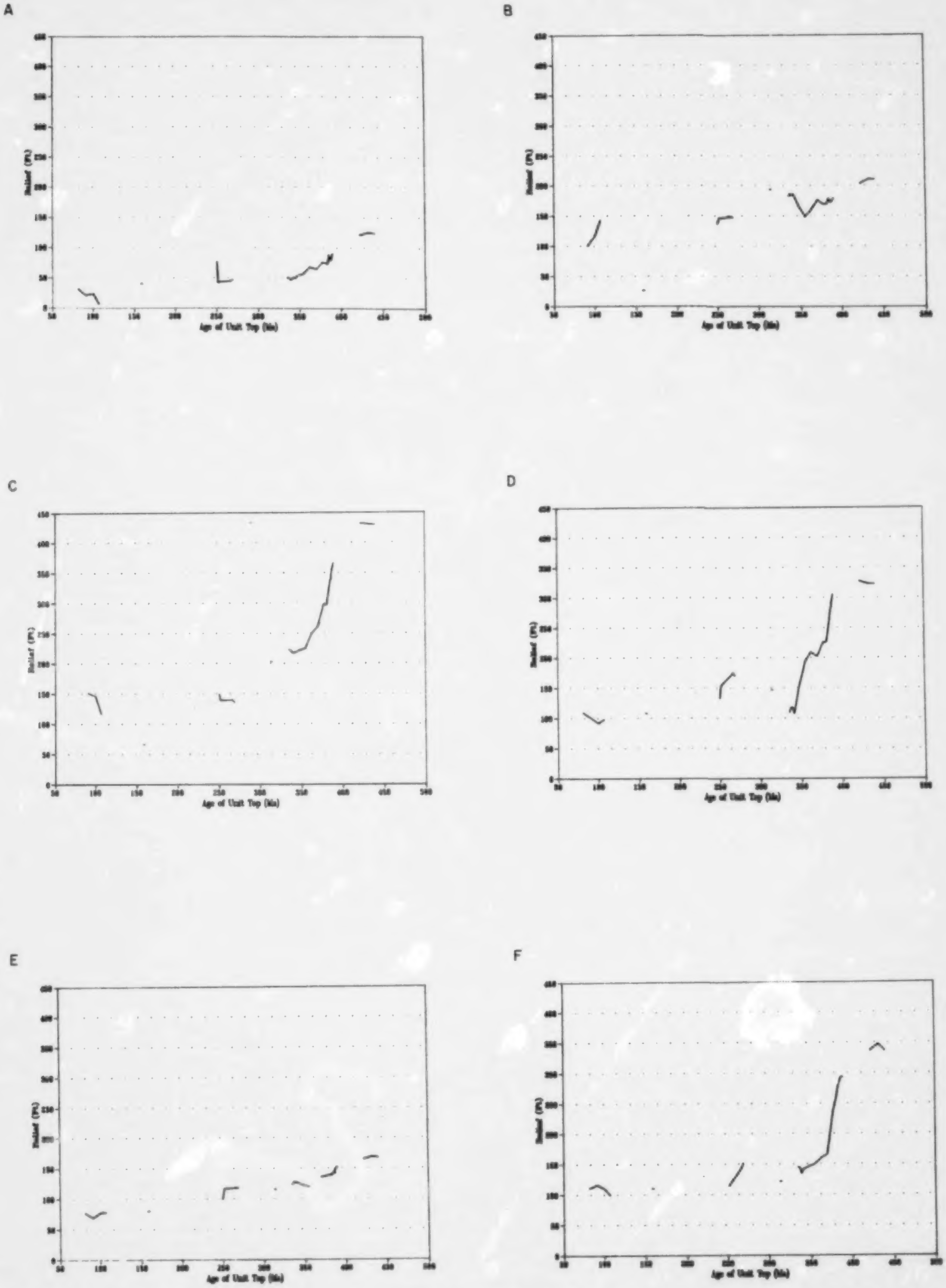

Figure 7, a) to f): Plots of structural relief. The absolute ages are from Ballard *et al.* (1983) and Harland *et al.* (1981): a) Boxcar Butte High: b) Rough Rider Anticline; c) Billings Anticline; d) Little Knife Anticline; e) Knutson Anticline; f) Richardton High.

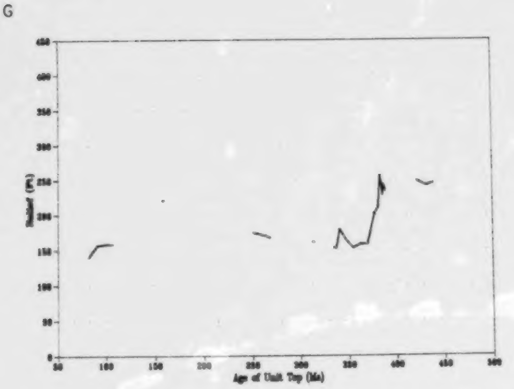

Figure 7 g): Taylor High

From the end of Lodgepole (355 Ma) to the end of the Pine Salt member of the Spearfish Formation, there was little change in relief, but net uplift of about 25 ft (8 m). Between Pine Salt and Rierdon Formation (160 Ma) about 20 ft (6 m) of uplift took place, and to the end of Niobrara (82 Ma), less than 10 ft (3 m).

Richardton High

The northwest-trending Richardton high is located under Richardton and Taylor fields in T139N, R92W - R93W.

There is about 80 ft (25 m) of relief on the Mowry, about 110 ft (34 m) on the Madison, 225 ft (70 m) on the Duperow, and about 300 ft (90 m) on the Red River Formation. The isopach maps show thinning of the Rierdon-Piper, Tyler, Kibbey, Charles-Mission Canyon, Three Forks, Duperow, Dawson Bay, Ashern, Interlake, and Stonewall Formations.

The plot of structural relief (Fig. 7f) indicates little relative uplift prior to the end of Stonewall time (422 Ma). Between Stonewall and the end of Ashern (389 Ma), about 45 ft (14 m) and between 389 Ma and 370 Ma (Three Forks Formation), about 130 ft (40 m) of uplift occurred. Then uplift continued at a slower rate, with only minor reversals, until sometime after the end of the Tyler (315 Ma), when another 45 ft (14 m) of uplift occurred. From the Tyler until the end of Opeche (269 Ma), about 30 ft (9 m) of subsidence took place. From Opeche to the end of Pine Salt Member (250 Ma), about 35 ft (11 m) of uplift occurred. Quiescence followed to the end of Inyan Kara (107 Ma). Between the Inyan Kara and Greenhorn (91 Ma), about 20 ft (6 m) of subsidence took place, followed by minor uplift by the end of Niobrara time (82 Ma).

Taylor High

The Taylor high is located in T140N, R93W, to the north of the town of Taylor, and trends approximately east-west.

There is about 100 ft (30 m) of relief across the high on the Mowry, 120 ft (37 m) on the Madison, 180 ft (55 m) on the Duperow, and about 220 ft (67 m) on the Red River Formation. Thinning over the high is evident on the isopach maps of the Niobrara-Carlile, Mowry, Tyler, Kibbey shale, Lodgepole, Three Forks, Birdbear, Duperow, and Interlake Formations.

The structural relief plot (Fig. 7g) indicates that very little uplift took place until after the end of the Dawson Bay (385 Ma). Between that time and the end of Three Forks (370 Ma), about 100 ft (30 m) of uplift occurred. There was no significant movement between Three Forks and the end of Kibbey time (335 Ma). Between Kibbey and the end of the Belfield Member of the Spearfish Formation (252 Ma), about 20 ft (6 m) of subsidence took place, and 15 ft (5 m) of uplift between end of Rierdon and end of Inyan Kara time (107 Ma). From the Inyan Kara until the end of Greenhorn (91 Ma), no significant movement occurred, but again 15 ft (5 m) of uplift between Greenhorn and the end of the Niobrara (82 Ma).

Stark County Low

In central and southern Stark County, an elongate low is evident on most of the structure contour maps above the Minnekahta Formation. The western end and deepest part of this feature is in T138N, R98W, where approximately 200 ft (60 m) of relief are present at most horizons. The feature extends across several townships to the northeast of T138N, R98W.

The Pine Salt Member of the Spearfish Formation is absent within the feature, but in adjacent wells more than 150 ft (45 m) are present. This suggests that the low is a result of dissolution of the Pine Salt, and is not directly due to tectonic activity. The timing of salt dissolution is uncertain. None of the wells in the low show a compensating thickening in any of the strata below the top of the Pierre Formation. Dissolution therefore appears to postdate Pierre deposition. The western end of the low is approximately coincident with the "South Heart syncline" (Caldwell, 1954), in which rocks of the White River Group (Oligocene) are deformed. If the two features are related, some or all of the solution collapse took place after the Oligocene.

SUBSIDENCE HISTORY

Subsidence history analysis was performed on all wells that penetrated to the Precambrian. Four of the subsidence history curves are shown in Figure 8. All of the curves are similar in form to the ones illustrated, except that the amount of calculated subsidence increases somewhat toward the basin center, and is slightly less for wells on structural highs.

All of the curves show the same pattern of subsidence. First, a sharp increase in rate at about 495 Ma, which probably represents the beginning of general basin subsidence. That event was followed by four others during the Paleozoic, each of which began with a high rate of subsidence, and declined to a relatively low rate. These events began: 1) at some time prior to 450 Ma, before the onset of Winnipeg deposition; 2) prior to the beginning of the Ashern deposition, about 400 Ma; 3) about 365 Ma, early in Lodgepole time; and 4) about the beginning of the Opeche (275 Ma). Very little movement took place from the end of the fourth event, about 225 Ma, until around the beginning of Pierre time (82 Ma), when rapid subsidence recurred. The duration of this event is unknown, as we have no reliable stratigraphic datum above the Pierre Formation. This pattern of subsidence seems to hold true for most of the North Dakota part of the Williston basin (LeFever, 1988).

Generally, the subsidence events are not coincident with major orogenic events elsewhere in North America, although they seem to reflect the response of the basin to activity in surrounding areas. The Devonian subsidence coincides with the opening of the Williston Basin to the Elk Point Basin, the Mississippian subsidence with activity in the Central Montana Trough, and the Late Cretaceous event with the onset of Laramide orogenic activity to the west.

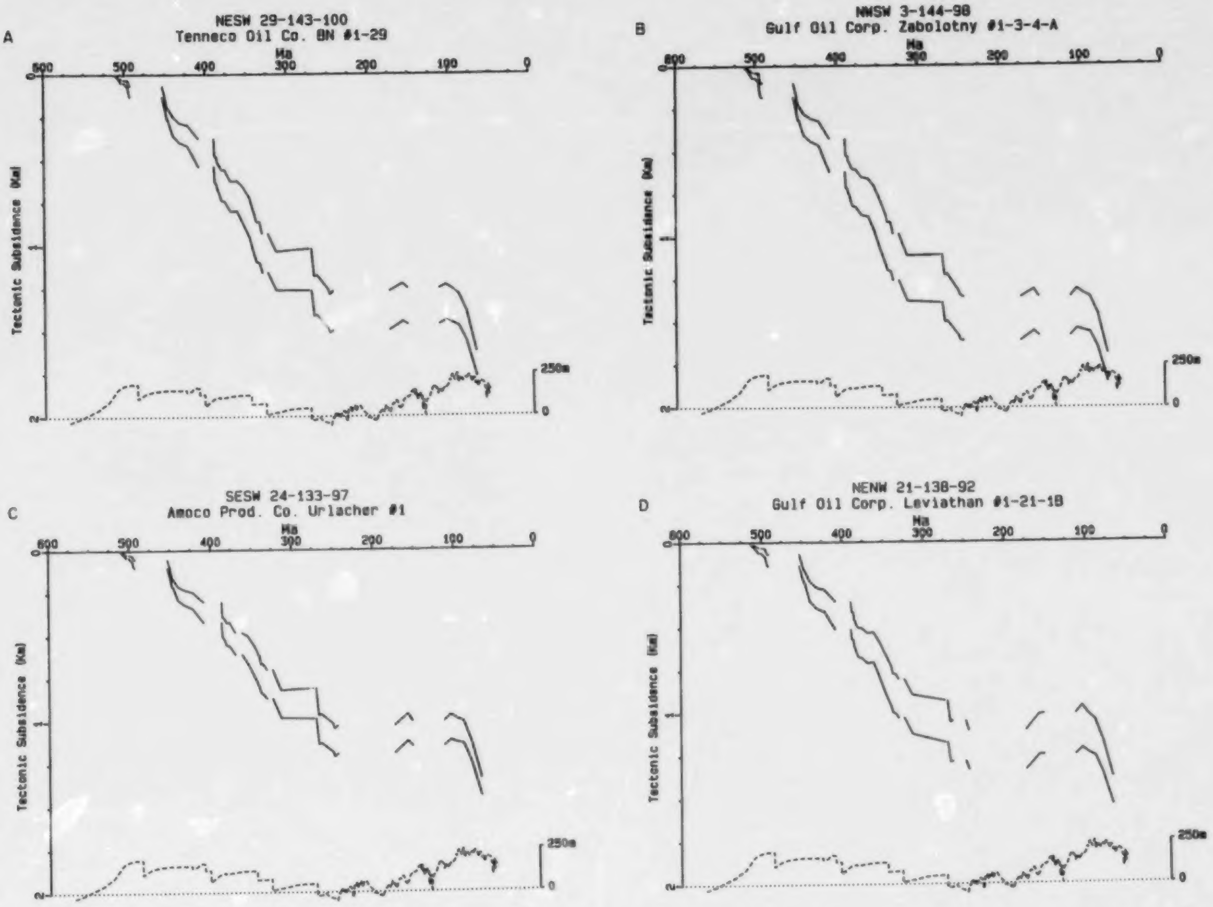

Figure 8: Subsidence history curves. Upper and lower curves are calculated subsidence for minimum and maximum compaction, respectively. Dashed line is Phanerozoic sea level curve, after Vail *et al.* (1977) and Haq *et al.* (1987), at the scale of the subsidence curves; the dotted line is present sea level. The absolute ages are from Ballard *et al.* (1983) and Harland *et al.* (1981).

SUMMARY

Several inferences can be drawn about development of the structural features within the area of study.

Most of the structural highs were generated during the Devonian and Mississippian, with the addition of some additional relief after Niobrara Formation deposition (82 Ma), presumably during Laramide activity. In some instances, it appears that some relief may have developed after deposition of the Stonewall Formation (422 Ma) and prior to the end of Ashern Formation deposition (389 Ma).

There are sufficient differences between the features in the isopach maps and the structural relief plots to indicate that all of the structures behaved independently of one another through most of the time interval documented.

The principal tectonic activity in the area was subsidence of the basin, the magnitude of which was considerably larger than differential movement of the structures. The structural features appear to be localities which responded to the basin movement, but which were intermittently decoupled from the general basin subsidence.

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GRAVITY MODELING OF PROTEROZOIC STRUCTURES IN NORTH-CENTRAL NORTH DAKOTA.

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ABSTRACT

During the Precambrian the core of the North American continent was formed by the collision and suturing of micro-continents. The resulting contact zone is exposed in Manitoba, and clearly marks the boundary between the Archean Superior Province and the Proterozoic Trans-Hudson Orogenic Belt. This zone includes the Thompson nickel belt and the Thompson Fault. Bouguer gravity maps of this area indicate that this boundary zone extends beneath the Phanerozoic cover in Manitoba and North Dakota.

In the study area, which is located between latitudes 48° and 49° N and longitudes 100° and 102°W, the westernmost major structure is the Thompson Fault. About 45 to 55 km (30 to 35 mi) east of this fault lies the north-south trending Superior - Trans-Hudson contact. Two-dimensional forward gravity modeling was used to determine the nature of the structures within the study area. Results indicate that (i) the Thompson Fault dips to the west and strikes about N13°E, running beneath the Minot area; (ii) the Superior-Trans-Hudson contact dips west and is slightly concave east; the dip steepens in the south; and (iii) a large circular positive gravity and magnetic anomaly centered near latitude 49° 15′N and longitude 101° W may represent a mafic pluton or a dense body of some other origin.

INTRODUCTION

During the Precambrian collision and suturing of micro-continents formed the core of the North American continent (Hoffman, 1990). Resulting structural and geochronological boundaries are present beneath the eastern flank of the Williston Basin in North Dakota and southern Canada (Fig. 1); Green *et al.*, 1979; Klasner and King, 1986). In the Williston Basin of western North Dakota the Precambrian basement is buried beneath 3 to 5 km (1.5 to 3 mi) of sedimentary rocks. Due to its great depth, the basement and its boundaries have been penetrated by only a few deep wells, and prior to this study had not been extensively modeled by geophysical methods such as gravity and magnetics.

This study explores the nature of the boundary between the Trans-Hudson orogenic belt and the Superior Province in north-central North Dakota, 48° to 49° N latitude, 100° to 102° W longitude (Fig. 1), using gravity modeling. This involves two-dimensional gravity modeling of four east-west profiles and one N77°W profile in the study area which includes portions of the Superior Province, the Thompson nickel belt and the Flin Flon-Snow Lake domain (Fig. 2). The approximate locations of geological features are taken from Green, Hajnal, and Weber (1985); gravity modeling has been done to extend the features into North Dakota and to help interpret the nature of possible structures at depth.

GEOLOGICAL SETTING

The Trans-Hudson orogen (Hoffman, 1981) is part of the Pan American orogenic system which extends from northern South Dakota, across Hudson Bay and east into Greenland and Labrador; it includes all the spatially associated Proterozoic belts (Lewry and Collerson, 1990). Hoffman (1988) suggested that the northern part of the Trans-Hudson orogen formed by collision between the Archean Superior and Hearn micro-continents during Precambrian time and that, to the south in the United States, collision between the Superior and Wyoming cratons formed the southern portion of the orogen (Fig. 1). Lewry *et al.* (1985) proposed that the Trans-Hudson orogenic belt extends as far west as the North American Central Plains (NACP) conductivity anomaly of Alabi *et al.* (1975) (Fig. 1). Oblique accretion of converging plates is suggested by a change in the direction of the Proterozoic polar wandering curve at about 1.65 Ga (Dutch, 1983).

The Trans-Hudson orogenic belt has been subdivided into several elements in Canada. These subdivisions were identified by field geology and geophysical characterization where they are exposed in Manitoba and Saskatchewan (Lewry *et al.*, 1985). Based on geophysical evidence, Green *et al.* (1979) and Klasner and King (1986) have predicted their presence in southern Canada and the northern United States. Rocks representing the eastern margin of the Trans-Hudson orogen crop out within the Churchill-Superior boundary zone (Stockwell, 1961; Lewry and Collerson, 1990) also known as the Archean-Proterozoic boundary zone (Green, Hajnal, and Weber, 1985;), the magnetic quiet zone (Klasner and King, 1986), the Thompson mobile belt (Machado *et al.*, 1987), and the ensialic belt of the Superior margin (Hoffman, 1988). The Thompson nickel belt forms the southern portion of this unit where it is exposed in Canada (Fig. 1). This name is used to refer to corresponding rocks in the study area. The term "Archean-Proterozoic" boundary zone will be used to include both the Thompson nickel belt and the Fox River belt (Fig. 1). A series of collisional events associated with this convergent plate boundary resulted in curvilinear belts of metamorphic and igneous rocks (Lewry *et al.*, 1985). The Thompson nickel belt may have been the site of Proterozoic crustal subduction (Green, Weber, and Hajnal, 1985) and a transcurrent fault (Thomas *et al.*, 1987). The Flin Flon-Snow Lake domain has geological characteristics which suggest that it may have formed as a volcanic island arc complex. Genetically related subduction ultimately wedged the Thompson nickel belt between the Superior block and the rest of the Trans-Hudson orogen (Green, Hajnal, and Weber, 1985).

GEOPHYSICAL EXPRESSION

The Archean-Proterozoic boundary zone generates one of the more prominent and continuous gravity anomalies in North America (Gibb, 1983; Fig. 3). This gravity signature is caused by

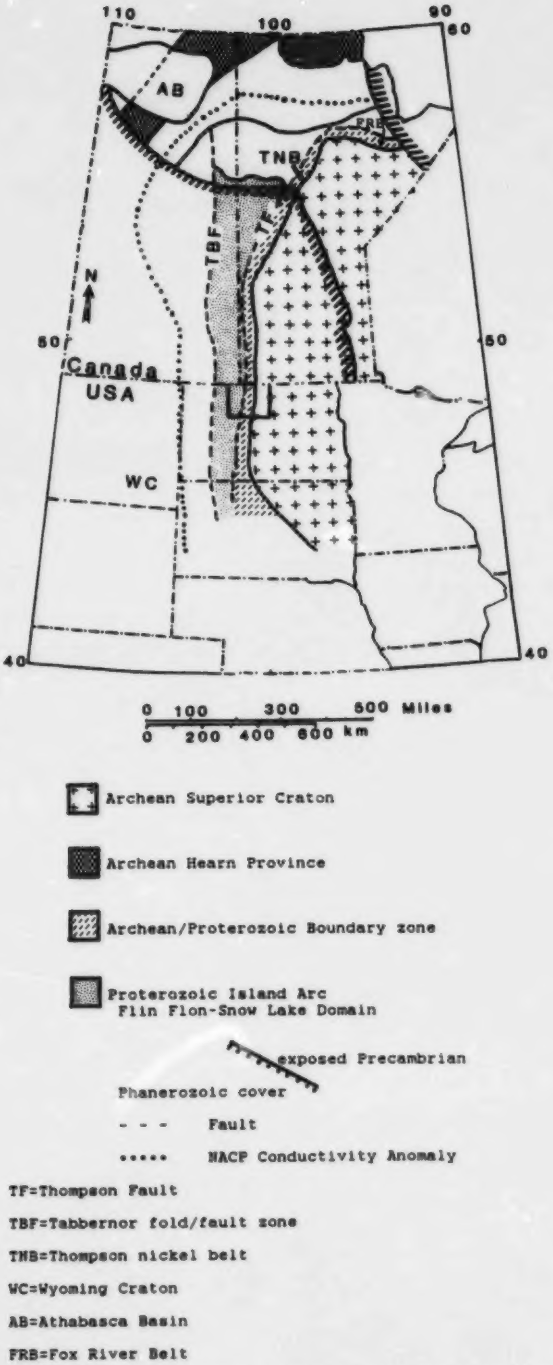

Figure 1: Simplified geological map of the Precambrian surface showing the location of the study area (rectangle in north-central North Dakota), with respect to the Archean Superior Province and a portion of the Proterozoic Trans-Hudson orogenic belt. Geological contacts are from Green, Hajnal, and Weber (1985). The location of the North American Central Plains (NACP) conductivity anomaly is taken from Alabi *et al.* (1975), Handa and Camfield (1984) and Gupta *et al.* (1985).

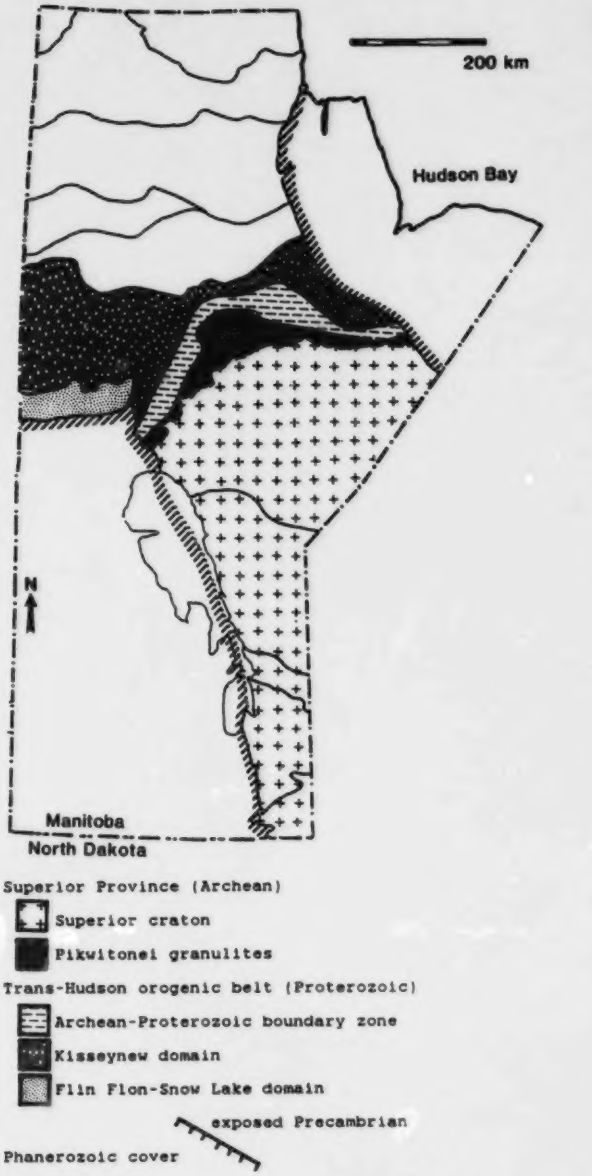
200 km
Hudson Bay
N
Manitoba
North Dakota
Superior Province (Archean)
Superior craton
Pikwitonei granulites
Trans-Hudson orogenic belt (Proterozoic)
Archean-Proterozoic boundary zone
Kisseynew domain
Flin Flon-Snow Lake domain
exposed Precambrian
Phanerozoic cover

Figure 2: Location of geological units discussed in text. Modified after Manitoba Department of Mines, Natural Resources and Environment (1979).

differences in structural fabric and general lithologies on either side of the suture (Gibb, 1983). The Superior Province is characterized by predominantly east-west gravity and magnetic fabrics (Fig. 4), whereas the Trans-Hudson is a region with more variable magnetic fabric and a northeast-southwest gravity trend (Bell, 1971; Green, *et al.*, 1979).

Geophysical data help to characterize the Trans-Hudson orogenic belt where it is exposed in Canada, and these characteristic gravity and magnetic signals delineate the belt underneath the Phanerozoic rocks of the Williston Basin. North-south trending geophysical signatures indicate that the Flin Flon-Snow Lake domain extends from its outcrop area southwards beneath Phanerozoic cover (Green, Weber, and Hajnal, 1985). The southern extension of the Thompson nickel belt has been inferred from continuation of a characteristic magnetic quiet zone which runs through the central part of the study area east of the presumed location of the Flin Flon-Snow Lake domain (Klasner and King, 1986). Geophysical characteristics typical of the Archean rocks of the Superior Province apparently extend into eastern North Dakota and northeastern South Dakota (Green, Weber, and Hajnal, 1985). The contact between the Superior craton and the Trans-Hudson orogenic belt (Sims and Peterman, 1986) within the study area parallels the western flank of a north-south, linear, 25 milligal positive gravity feature (Lyons and O'Hara, 1982). The Thompson Fault, which defines the western side of the Thompson nickel belt (Fig. 1), lies within the study area (Green, Hajnal, and Weber, 1985) and has been detected in modeling of seismic profiles in southern Manitoba, 50 km north of the study area (Morel-a-l'Huissier *et al.*, 1987).

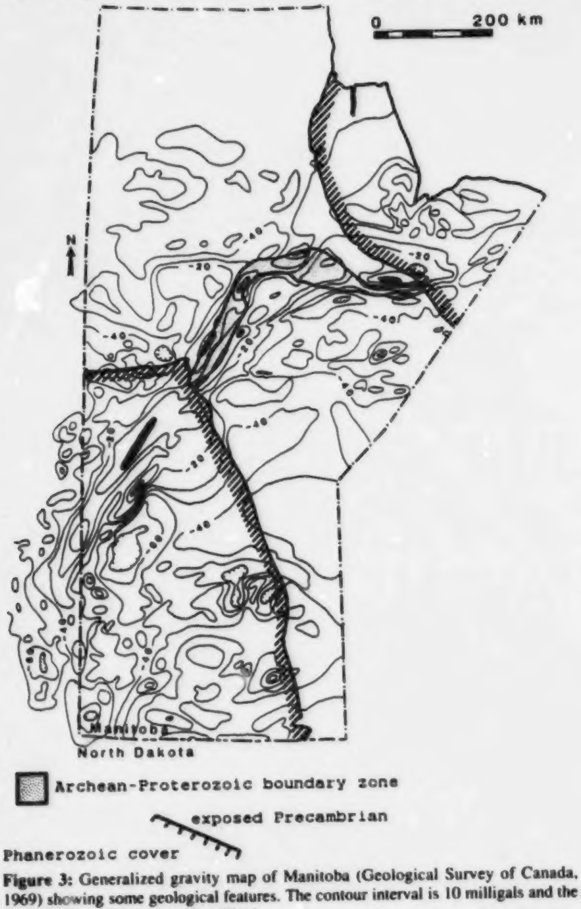

Figure 3: Generalized gravity map of Manitoba (Geological Survey of Canada, 1969) showing some geological features. The contour interval is 10 milligals and the solid black areas are 0 to +20 milligals.

LITHOLOGIES AND AGE DATES

The Precambrian basement in the study area underlies Phanerozoic sedimentary rocks of the Williston Basin; depth of burial ranges from about 1750 to 4000 m (1 to 2.5 mi). Few wells penetrate completely the Phanerozoic cover so lithological details of the Precambrian surface are poorly known. The Precambrian structure map of the study area reveals the northeasterly flank of the Williston Basin (Fig. 5). Large scale topographic features were apparently eroded prior to deposition of the Cambrian Deadwood Formation. Karner *et al.* (unpublished manuscript, 1981) and Lidiak (unpublished manuscript) have studied Precambrian core samples and cuttings from wells drilled in North Dakota. Carlson (unpublished table) also compiled a list of lithologies. These authors are in some disagreement about rock-type identification for many of these wells.

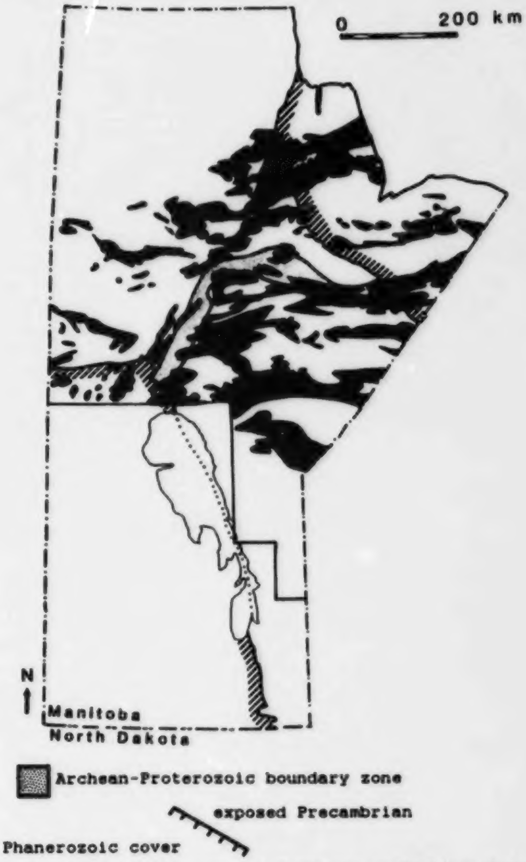

Figure 4: Generalized magnetic map of Manitoba, modified from Morley *et al.* (1967). The solid areas are greater than +200 gammas.

In general, Archean radiometric dates have been obtained east of the study area (Burwash *et al.*, 1962; Peterman and Hedge, 1964) and Proterozoic dates to the west (Peterman and Hedge, 1964). Within the study area, the data show that both Archean and Proterozoic supracrustal rocks were deformed, metamorphosed, and intruded by granite during the Hudsonian orogeny (Peterman and Goldich, 1982).

METHOD OF STUDY

BOUGUER GRAVITY

The gravity data used in this study are available on North Dakota Geological Survey tape #S4395, compiled by the United States Department of Defense. On a regional Bouguer gravity map, anomalies due to large-scale, deep-seated structures predominate; therefore, some reduction has been done to resolve anomalies of interest. The object of this research was to investigate structural relationships in the Precambrian rocks, so procedures were applied to remove the gravity field caused by the Phanerozoic rocks of the Williston Basin from the observed gravity. Figure 6 presents the Bouguer gravity map before the effect of the Phanerozoic mass was removed. Bouguer gravity was recalculated to a new datum at 3450 m (11,319 ft) below sea level, that is, below the Phanerozoic-Precambrian contact (Fig. 7).

A 415 milligal constant was added to each value to normalize each profile to a zero base in order to simplify the use of the modeling program. Because structures delineating the eastern margin of the Trans-Hudson orogen are approximately linear, they can be solved for by using a two-dimensional form of analysis (Talwani *et al.*, 1959).

ASSUMPTIONS MADE FOR MODELING

The accuracy of modeling depends on the applicability of the assumptions used. The specific assumptions used here are:

1) that geological bodies can be approximated by two-dimensional polygonal masses of uniform density; Talwani *et al.* (1959) point out that this analogy is adequate for large-scale linear features such as continental margins, for which an infinite strike can be assumed, although less appropriate for small non-linear bodies; a significant non-linear anomaly is present in the study area and is therefore modeled by the method discussed below;
2) that the Thompson Fault and the Superior-Trans-Hudson contact are located within the study area where Green, Hajnal, and Weber (1985) have drawn them;
3) that the earth's crustal thickness is between 41 and 45 km (25 and 28 mi) (Morel-a-l'Huissier *et al.*, 1987);
4) that the tectonism in the area included a convergent plate boundary due to a continent - island arc collision (Green, Hajnal, and Weber, 1985);
5) that current plate tectonic theory applies to the Proterozoic earth;
6) that the density of crustal rocks increases with depth within the polygons and in the non-enclosed area surrounding the polygons; and
7) that the existing crustal structure is far more heterogeneous and complex than can be resolved using gravity data alone.

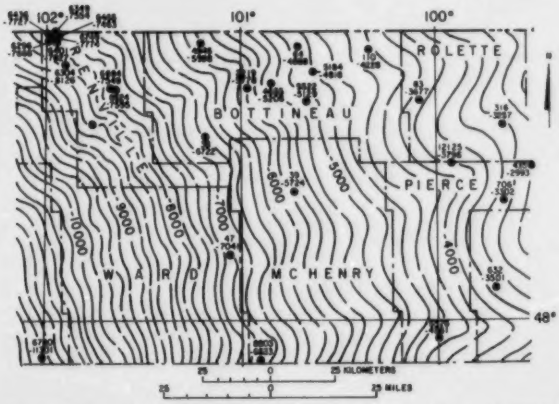

Figure 5: Precambrian structure map (Heck, 1988). The contour interval is 200 ft (61 m) and datum is sea level. Reproduced with permission of the North Dakota Geological Survey.

MODELING

Four east-west gravity profiles, A, B, C, and D, were constructed from selected data points shown on Figure 8. These profiles are oriented east-west because structures defining the Superior-Trans-Hudson boundary in the study area are linear and

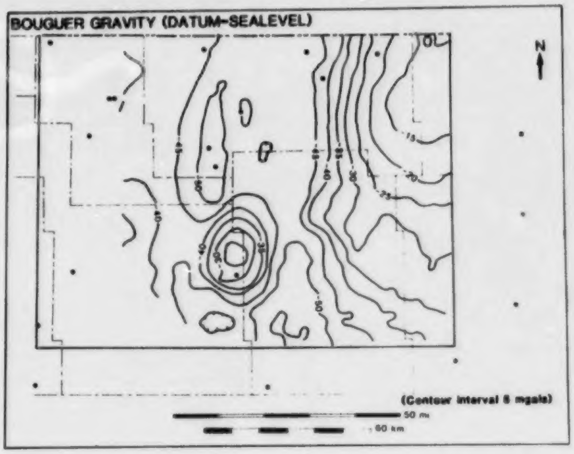

Figure 6: Bouguer gravity map. Sea level is datum. Contoured by the Surface II graphics system (Sampson, 1978). Solid circles represent Precambrian test wells.

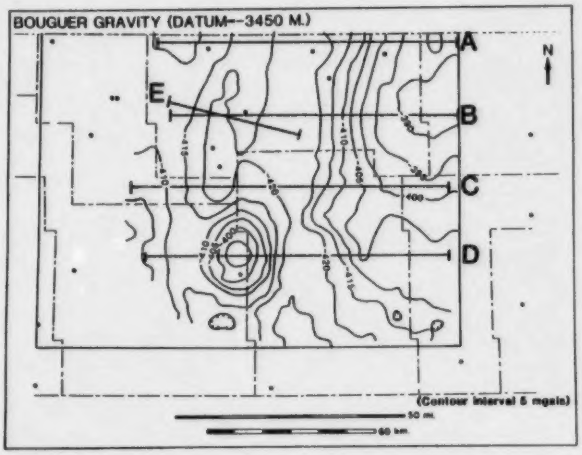

Figure 7: Bouguer gravity map showing locations of Profiles A, B, C, D, and E. Datum is 3450 m (11,319 ft) below sea level. Contoured by the Surface II graphics system (Sampson, 1978). Solid circles represent Precambrian test wells.

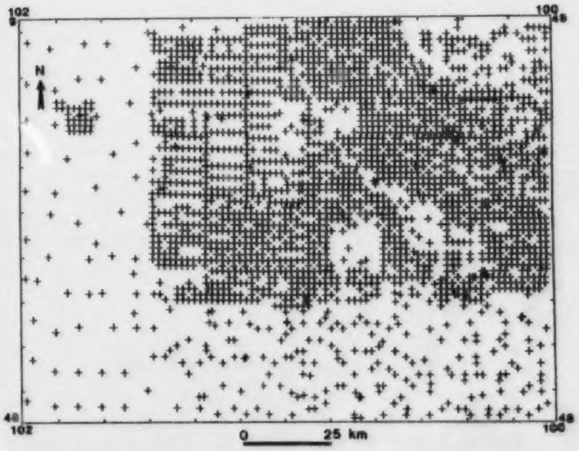

Figure 8: Gravity station locations within the study area. The boundaries of the study area are latitudes 48° and 49° N and longitudes 100° and 102°W.

north-south in orientation (Green, Hajnal, and Weber, 1985; Klasner and King, 1986). Contouring and profiles are limited to the area between latitudes 48° 15′ and 49° N, and longitudes 100° and 101° 15′ W because data are closely spaced (Fig. 8). The profiles are located about 25 km (15 mi) apart. Profile D bisects a circular gravity high in the south-central part of the study area (Fig. 7), an anomaly coincident with a large, significant magnetic high (Zietz, 1982). A fifth profile, E (Fig. 7), was constructed and modeled after the first four showed that the Thompson Fault strikes approximately N13°E.

Lithological density contrasts were based on data from Telford *et al.* (1976) and measured densities of Canadian rocks by Gibb (1968). Rock types in North Dakota are assumed to be essentially the same as in Manitoba so densities used for modeling were based, in part, on the following petrological information. In Manitoba, immediately east of the Thompson nickel belt, the Kenoran (2.5 to 3.1 Ga) granitoid-gneiss terrane of the Pikwitonei region granulites represents a lower crustal equivalent of east-west-trending Archean granite-greenstone belts (Green, Weber, and Hajnal, 1985; Fig. 2). The Thompson nickel belt consists of metavolcanic and metasedimentary rocks of granulite facies, overprinted by amphibolite facies metamorphism (Weber and Scoates, 1978). Many lenticular and sill-like ultramafic bodies of serpentinized peridotite occur in the supracrustal rocks and basement gneiss (Peredery *et al.*, 1982). The north-south trending Flin Flon-Snow Lake domain consists of a thick sequence of low-potassium tholeiitic to calc-alkaline basalts and andesites, pyroclastics and subordinate rhyodacites rocks (Green, Hajnal, and Weber, 1985).

Profiles A, B, C D and E (Fig. 7) were modeled using a compiled computer program (Malinconico, 1987) based on work reported by Talwani *et al.* (1959). The program calculates the gravity signal attributed to each body and sums the total signal. The gravity curve obtained from the polygonal model is compared to the observed curve, and changes are made to get the closest fit between the assumed geology, observed data and modeled gravity values.

A second method was used to model the large circular anomaly crossed by Profile D (Fig. 7). The gravity signal of this feature could not be calculated using the method of Talwani *et al.* (1959) because it is not a linear feature. This body was therefore assumed to be similar to a cone and vertical cylinder. This anomaly was modeled using a BASIC language computer program (Braun, 1987) which calculated the gravity due to a body by approximating its shape to a stack of discs. The gravity (g) of one disc in the stack at point P on the horizontal plane (Fig. 9) is given by

$$g = G\sigma\pi \frac{a^2h^2}{(r^2 + z^2)}$$

where G=the universal gravitational constant, σ=the density contrast of the disc, a=the radius of the disc, h=the height of the disc, z=depth to the centre of the disc, and r=the distance between P and the centre of the disc. The gravity for all the discs in the stack is then summed to obtain the total signal. The profile generated for this body alone is then added to the curve calculated for the rest of the polygonal model, and the result compared to the observed curve.

RESULTS AND ANALYSIS

PHANEROZOIC-PRECAMBRIAN INTERFACE

At this scale, the Williston Basin contributes little to the shape and amplitude of the Bouguer anomalies in the study area. The absolute value of corrected Bouguer gravity (Fig. 7) is about 380 mgals less than gravity at sea level (Fig. 6). This is because the Bouguer gravity has been uniformly over-corrected for the area between sea level and -3450 m (11,319 ft). For modeling these anomalies, the Precambrian basement, and not the Phanerozoic overburden, is therefore the source of the residual gravity anomalies in the study area (Fig. 7).

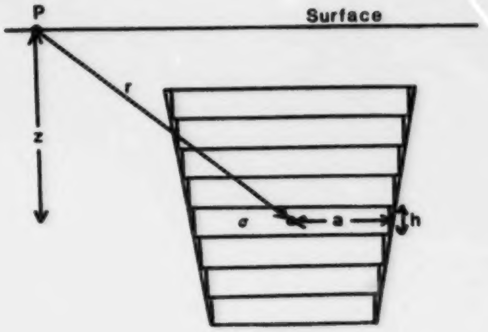

Figure 9: Geometric representation of the variables used to calculate the gravitational attraction of a stack of graduated discs (Braun, 1987); σ=density contrast of the disc, a=radius of the disc, h=height of the disc, z=depth to the centre of the disc, and r=distance from P to the centre of the disc.

GRAVITY MODELS

Polygonal gravity models designed for this study area are based on previous work, available lithologies, age dates, and a nearby seismic model. Profiles A, B, C, and D and corresponding models are presented in Figures 10 to 13 respectively. The locations of the profiles are shown on Figure 7.

Profile A

The polygonal gravity model for profile A (Fig. 10) has three main compartments and a smaller one. The easternmost, large polygon represents all rocks east of the contact between the boundary zone (the central polygon) and the Superior Province. Lithologies for this area are mostly granites and granodiorites. No age dates have been collected in this part of the study area, although several Archean ages are recorded in eastern North Dakota (Burwash *et al.*, 1962). The upper portion of the eastern polygon has a density contrast of .035 g cm^{-3}, and the lower portion of .04 g cm^{-3} with respect to the surrounding non-enclosed area. These densities are in the range of granodiorite, granulite, phyllite, and gneiss. The contact between this polygon and the central polygon dips to the west.

The central polygon represents the rocks beneath the Thompson nickel belt. Basement surface lithologies are generally schists and granites in this area. The upper portion of the central polygon has a density contrast of -.025 g cm^{-3}, and the lower portion -.03g cm^{-3}. These densities are in the range of granite, andesite, schist, and metagreywacke. The western side of the central polygon corresponds to the west-dipping Thompson Fault (Morel-a-l'Huissier, 1987; Fig. 1). The western polygon represents the Hudsonian rocks to the west of the Thompson Fault. Rocks identified in this area are massive granodiorite and gneiss for which an early Proterozoic age (1.76 Ga.) has been reported (Peterman and Goldich, 1982). The upper portion of this body has a density contrast of .015 g cm^{-3} and the lower portion has a density contrast of .02 g cm^{-3}. These densities are in the range of gneiss, phyllite, granulite, greywacke, and quartzite.

A small polygon within the easternmost body, situated at the basement surface, has a density contrast of -.08 g cm^{-3}, in the density range for granite. Another small polygon within the central polygon has a density contrast of .15 g cm^{-3}, in the density range of greenstones.

Profile B

The polygonal gravity model for Profile B (Fig. 11) also features three large compartments and one small one. The easternmost polygon represents all rocks to the east of the Superior-Thompson nickel belt contact, where the main lithologies are massive granodiorite, biotite granite, and granite. The density contrast of the upper portion of this body is .035 g cm^{-3}, that of the lower portion, .04 g cm^{-3} (granodiorite, granulite, gneiss, phyllite). The contact between the eastern and central bodies dips west, and is slightly steeper than in Profile A.

The large central polygon represents the Thompson nickel belt. Two wells penetrate basement which is reportedly made up of amphibolite, biotite-epidote-quartz-hornblende-andesine schist, serpentinite, granite, hornblende schist, and chlorite schist. The density contrast of the upper portion of this body is -.03 g cm^{-3} and of the lower portion is .025 g cm^{-3} (granite, andesite, schist, or metagreywacke). The western side of this polygon represents the west-dipping Thompson Fault (Morel-a-l'Huissier *et al.*, 1987). The large western polygon represents Hudsonian rocks to the west of the Thompson Fault. The density contrast in the upper portion of this body is .015 g cm^{-3} and in the lower portion is .02 g cm^{-3} (gneiss, phyllite, granulite). A small polygon at or near the basement surface has a density contrast of .15 g cm^{-3} (greenstone). Two age dates from near the western end of Profile B in the general vicinity of the western most body show basement rocks in NDGS well 6624 to be Archean (2.9 Ga) and in NDGS well 6684 to be Proterozoic (1.9 Ga) (Peterman and Goldich, 1982).

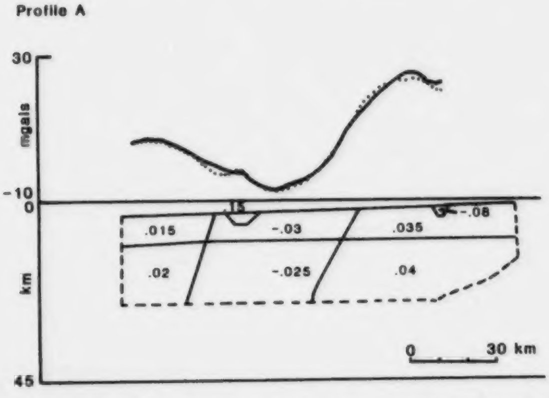

Figure 10: Polygonal gravity model for Profile A. The vertical scale for the top half is -10 to 30 mgals. The vertical scale for the bottom half is 0 to 45 km (0 to 30 mi). The horizontal scale is in km. The solid curve represents observed gravity and the dotted curve represents gravity calculated from the model. The polygons are labeled with density contrast in g cm^{-3}.

Profile C

The polygonal gravity model for profile C (Fig. 12) has three large compartments and two small. The large eastern polygon again represents the rocks (mainly granite, biotite granite and granodiorite) east of the Superior-Thompson nickel belt contact. The density contrast of the upper portion of this body is .025 g cm^{-3} and of the lower portion, .04 g cm^{-3} (granodiorite, granulite, phyllite, or gneiss). The contact between this body and the large central polygon dips west and is concave east.

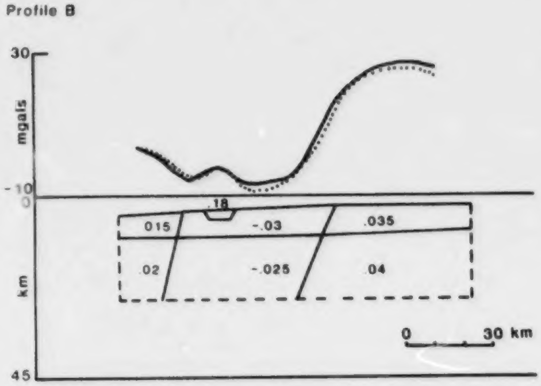

Figure 11: Polygonal gravity model for Profile B. The vertical scale for the top half is -10 to 30 mgals. The vertical scale for the bottom half is 0 to 45 km (0 to 30 mi). The horizontal scale is in km. The solid curve represents observed gravity and the dotted curve represents gravity calculated from the model. The polygons are labeled with density contrast in g cm^{-3}.

The central polygon represents the rocks of the Thompson nickel belt, generally granite and chlorite schist. The upper portion of the central polygon has a density contrast of -.03 g cm^{-3}, the lower portion, -.025 g cm^{-3} (granite, andesite, schist, or metagreywacke). The western boundary of this polygon represents the west-dipping Thompson Fault (Morel-a-l'Huissier *et al.*, 1987).

The western polygon represents Hudsonian rocks, reportedly gneisses, west of the Thompson Fault. The upper portion of this body has a density contrast of .02 g cm^{-3}, the lower portion, .025 g cm^{-3} (gneiss, phyllite, or granulite).

A small polygon in this model, located at or near the basement surface, has a density contrast of .14 g cm^{-3} (greenstone).

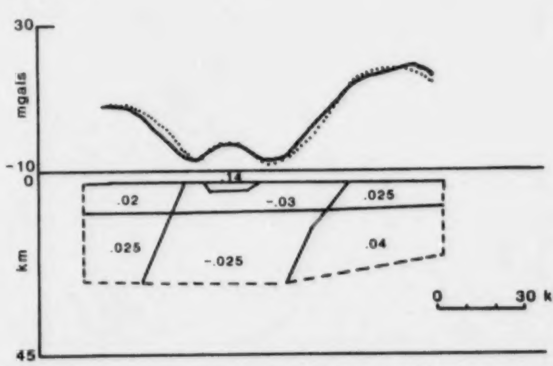

Figure 12: Polygonal gravity model for Profile C. The vertical scale for the top half is -10 to 30 mgals. The vertical scale for the bottom half is 0 to 45 km (0 to 30 mi). The horizontal scale is in km. The solid curve represents observed gravity and the dotted curve represents gravity calculated from the model. The polygons are labeled with density contrast in g cm^{-3}.

Profile D

The polygonal gravity model for Profile D (Fig. 13) comprises three large compartments and several smaller ones. The large eastern polygon represents the Superior Province, largely made up of granodiorite and granite. The density contrast of the upper portion of the body is .02 g cm^{-3}, of the lower portion, .035 g cm^{-3} (granodiorite, granulite, phyllite, or gneiss). The contact between this polygon and the central polygon dips west and is steeper here

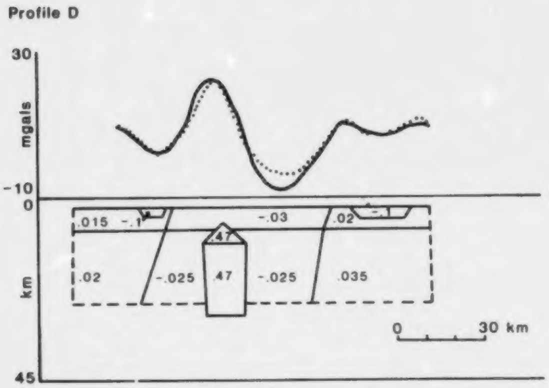

Figure 13: Polygonal gravity model for Profile D. The vertical scale for the top half is -10 to 30 mgals. The vertical scale for the bottom half is 0 to 45 km (0 to 30 mi). The horizontal scale is in km. The solid curve represents observed gravity and the dotted curve represents gravity calculated from the model. The polygons are labeled with density contrast in g cm^{-3}.

than to the north. The upper portion of the central polygon has a density contrast of -.025 g cm^{-3}, the lower portion, -.03 g cm^{-3} (granite, andesite, schist, or metagreywacke). The western side of this polygon represents the west-dipping Thompson Fault (Morel-a-l'Huissier *et al.*, 1987). The large western polygon features Hudsonian rocks to the west of the Thompson Fault. The upper portion of this body has a density contrast of .015 g cm^{-3} and the lower portion .02 g cm^{-3} (gneiss, phyllite, greywacke, or granulite).

Shallow bodies in each of the eastern and western polygons are assigned a density contrast of -.1 g cm^{-3} (granite). Within the large central polygon, a large bullet-shaped form has been modeled in two parts to account for the large circular gravity high (Fig. 13). The top was modeled as a cone, the bottom as a tapered cylinder. The cone has a bottom radius of 7,000 m (23,000 ft), a top radius of 300 m (1,000 ft), is 4,000 m (13,000 ft) high and the top is 7,000 m (23,000 ft) below sea level. The tapered cylinder has a bottom radius of 6,400 m (21,000 ft), a top radius of 7,000 m (23,000 ft) and a height of 17,000 m (55,000 ft). Both parts of the body have a density contrast of .47 g cm^{-3} (gabbro, peridotite, or pyroxenite).

From the profiles, the boundary between the Superior Province and the Thompson nickel belt and the Thompson Fault was plotted on the Bouguer gravity map (Fig. 14). From each profile the top of the upper polygon (T), the top of the lower polygon (M), and the bottom of the lower polygon (B) were plotted on the map for both of the major structures. This map shows that the Superior-Thompson nickel belt contact is approximately linear (north-south) perpendicular to the profiles. The Thompson Fault is also approximately linear; contrary to previous work (Green, Hajnal, and Weber, 1985), this fault strikes about N13°E, and is therefore not perpendicular to the traces of the profiles. Because the method of Talwani *et al.* (1959) requires that the polygonal bodies be infinite in extent perpendicular to the profiles, a fifth profile, profile E, was constructed and modeled perpendicular to the trace of the fault.

Profile E

The polygonal gravity model for profile E (Fig. 15) has two large compartments and one smaller. The large eastern body represents the rocks of the Thompson nickel belt. The upper portion of the eastern compartment has a density contrast of -.03 g cm^{-3}, the lower portion, .025 g cm^{-3} (andesite, granite, schist, or metagreywacke). The large western compartment represents Hudsonian rocks west of the Thompson Fault. The upper portion has a density contrast of .015 g cm^{-3}, the lower .02 g cm^{-3} (gneiss, phyllite, greywacke, or granulite). The small polygon in this model is at or near the basement surface, and has a density contrast of .14 g cm^{-3} (greenstone).

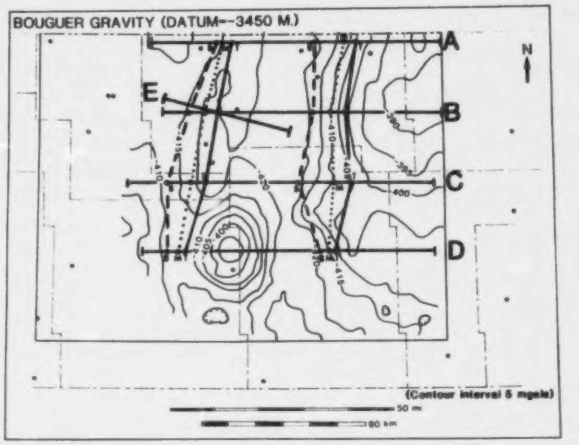

Figure 14: Bouguer gravity map showing the location of profiles A, B, C, D and E. The location of the Superior-Thompson nickel belt contact and of the Thompson Fault are shown as predicted from modeling.

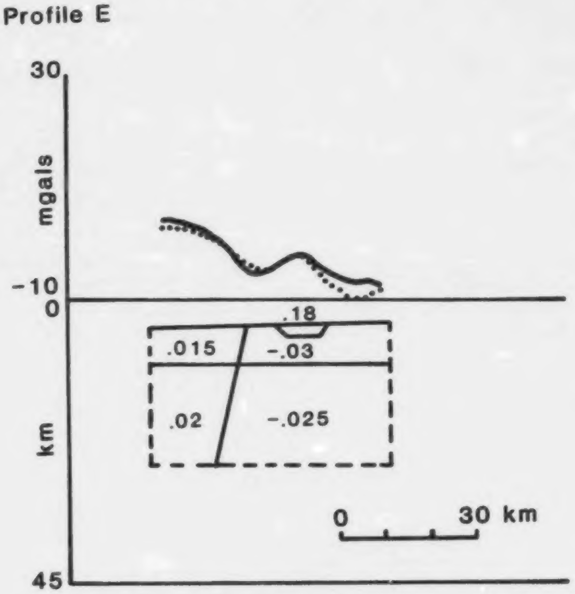

Figure 15: Polygonal gravity model for Profile E. The vertical scale for the top half is -10 to 30 mgals. The vertical scale for the bottom half is 0 to 45 km (0 to 30 mi). The horizontal scale is in km. The solid curve represents observed gravity and the dotted curve represents gravity calculated from the model. The polygons are labeled with density contrast in g cm^{-3}.

ANALYSIS

All five profiles reveal a similar basement structure with respect to the Superior-Thompson nickel belt contact and the Thompson Fault. The Superior-Thompson nickel belt contact dips to the west in each case and tends to steepen to the south. In Canada, 50 km (31 mi) north of the study area, the Thompson Fault is shown in seismic models (Morel-`a-l'Huissier *et al.*,

1987) to dip westwards with similar angle and depth of penetration to those in the gravity model.

In profiles A and D, the small polygons within the easternmost polygon may represent granitic bodies within the granite greenstone belts of the Superior craton. In profiles B and C, similar polygons to the west of the Superior boundary zone contact may represent allochthonous greenstone bodies from the Superior craton or be of another origin and emplaced during a continent-island arc collision (Lewry *et al.*, 1985). The small polygons above the eastern and western polygons in profile D may represent granitic bodies at or near the basement surface.

The circular positive anomaly crossed by Profile D is modeled as a deep bullet shaped body, perhaps a mafic mass. The density of this body falls within the average range for gabbro, peridotite, and pyroxenite. This gravity high coincides with a prominent magnetic high (Zietz, 1982).

Data on basement lithologies and age dates are sparse in the study area and possibly provide misleading information in this type of study. Lidiak *et al.* (1985) have pointed out that drill cuttings used in any study may have been collected from nonrepresentative sites or may have missed important anomalous bodies. This project focuses on the deep basement, but all sampling was from in the first few metres of Precambrian basement. No direct evidence currently exists for the actual age or lithology of deep basement rocks. Because of this and because most age dates and lithological information are distant from the profiles, the data were used only in the most general way in the development of the models.

Studies of seismic refraction and deep crustal reflection have revealed that the earth's crust is heterogeneous both laterally and vertically (Mueller, 1977; Oliver and Kaufman, 1977; Smithson and Shive, 1977). In this area, the basement is metamorphic and is probably discontinuously layered on many scales. Complex basement structures cannot be uniquely resolved from gravity data alone. Consequently, only the most general structures have been modeled. Seismic refraction modeling done in Canada, 50 km north of the study area, shows abrupt velocity changes at 10 km and between 30 and 40 km depth near the Thompson Fault (Morel-`a-l'Huissier *et al.*, 1987). In another seismic study in the Thompson nickel belt, distinct velocity changes were inferred at 2 km, between 11 and 13 km, and between 35 and 40 km (Green *et al.*, 1980). The models in this study are partially based on the seismic models of Morel-`a-l'Huissier *et al.* (1987). Polygons in the models presented here generally show a density change at a depth of about 10 km and do not extend deeper than 30 km.

Preliminary results of recent U-Pb geochronology in Canada indicate that the Thompson nickel belt was deformed by the Hudsonian event between 1.884 and 1.786 Ga (Machado *et al.*, 1987). Van Shmus and Bickford (1987) obtained zircon dates between 1.89 and 1.83 Ga for all sampled rocks from the Trans-Hudson orogenic belt in Manitoba and Saskatchewan. In this region of Canada, subduction related collision between island arc and continent apparently took place in early Proterozoic times (Lewry, 1981; Green, Weber, and Hajnal, 1985). The Thompson nickel belt is inferred to be the site of subduction and a transcurrent fault. To the west, the Flin Flon-Snow Lake domain represents an island arc complex which formed during crustal subduction and wedged the Thompson nickel belt between the Superior Province and the rest of the Trans-Hudson belt (Green, Hajnal, and Weber, 1985).

Green, Hajnal, and Weber (1985), Green, Weber, and Hajnal (1985), and Klasner and King (1987) contend that an ocean existed to the west and the north of the Superior Craton (Fig. 16A). Subduction of oceanic crust (Fig. 16B) generated an island arc, now the Flin Flon-Snow Lake domain (Fig. 16C). A sedimentary wedge formed in the resultant trench and was fed by continent and ocean sediments to the east and by volcanic sources to the west (Fig. 16C). Eventually, collision (Fig. 16D) of the continent with the island arc juxtaposed the Superior craton, the sedimentary wedge and the island arc, deforming portions of all three elements. Following collision and suturing, transcurrent faulting, perhaps caused by oblique plate interactions during the Hudsonian orogeny, occurred on the west side of the Thompson nickel belt (Green, Hajnal, and Weber, 1985). Uplift and erosion, possibly to the level of the dashed line in Figure 16D, followed by Phanerozoic downwarp and sedimentary infilling of the Williston Basin, would have resulted in the present apparent configuration (Fig. 16E). The models presented in this study provide supporting evidence for the plate collision models inferred from surface geology.

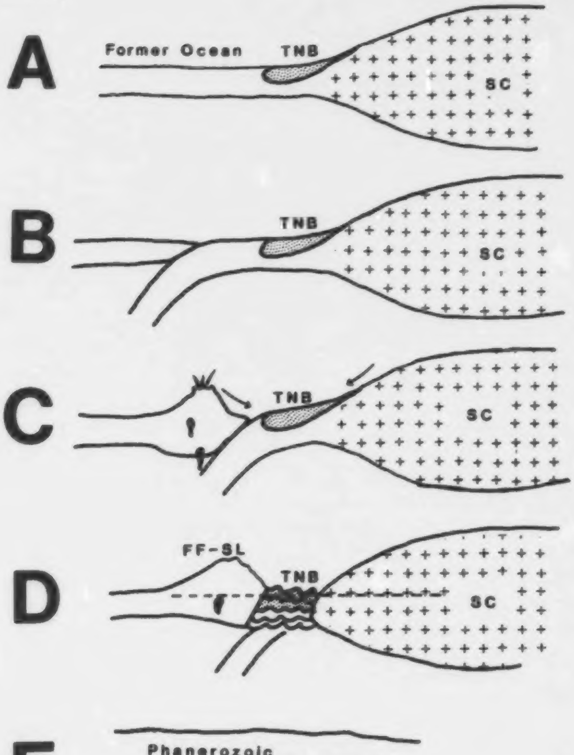

Figure 16: Crustal evolution of the boundary zone. A, B, C, and D are based on Green, Hajnal, and Weber (1985) and Green, Weber, and Hajnal (1985). A: Former ocean, possibly formed by rifting of the Archean crust. The Thompson nickel belt (TNB) formed as a result of continental margin development. B: Initiation of subduction. C: Island arc formation and trench sedimentation. D: Collision and suturing, followed by transcurrent faulting on the west side of the TNB. E: Final configuration after uplift, erosion, downwarp, and sedimentation.
SC= Superior Craton
FF-SL= Flin Flon-Snow Lake domain
- - - - possible erosion level

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ROLE OF SALT DISSOLUTION IN CONTROLLING OUTCROP DISTRIBUTION IN SOUTH-CENTRAL SASKATCHEWAN

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ABSTRACT

Dissolution of the Middle Devonian Prairie Evaporite is a major structure-forming process in the south-central Saskatchewan segment of the Williston Basin. Dissolution occurred as a multistage process of salt removal which included eventual collapse of the overlying sedimentary column and accelerated sedimentation into surface depressions produced during collapse. The dissolution process began soon after deposition of the Middle Devonian salts and has continued to the present. Salt removal of late Mesozoic and Cenozoic age, interpreted from post-Second White Speckled Shale (Upper Cretaceous) dissolution trends, is apparent at several locales in the area of investigation (Saskatchewan Twp. 1-17, Rge. 15-30 W2M; Montana Twp. 36N-37N, Rge. 44E-58E; North Dakota Twp. 162N-164N, Rge. 102W-103W). The late-stage solution collapse features, coupled with the erosional effects of Pleistocene glaciation, played a significant role in producing the current outcrop distribution in south-central Saskatchewan.

INTRODUCTION

The process of subsurface dissolution of Mississippian and Devonian evaporites in the Williston Basin area has been known for more than 30 years. Sloss and Laird (1947) and Baillie (1953) described the presence of evaporite solution breccias in the Upper Devonian of Montana and the Middle Devonian of Manitoba and Saskatchewan. The effect of subsurface salt removal on the structural evolution of younger strata was first noted by Bishop (1954). Subsequent written accounts include those of DeMille *et al.* (1964), Smith and Pullen (1967), Swenson (1967), Holter (1969), Gendzwill (1978), McTavish and Vigrass (1987) and McTavish (1990).

Dissolution features similar to those observed in the Williston Basin occur in many sedimentary basins. The phenomenon has been described in Alberta (Oliver and Cowper, 1983; Hopkins, 1987), Texas (Baumgardner *et al.*, 1982), the North Sea (Lohmann, 1972; Jenyon, 1988), the Mediterranean (Camerlenghi and Ceta, 1987), and the Sergipi - Alagoas continental margin basin of northeastern Brazil (Vigrass, *pers. comm.*). Jenyon (1986) gives worldwide examples with a clear description of the salt removal process.

The area of investigation (Fig. 1) covers 28,000 km^2 (11,000 miles^2) of south-central Saskatchewan (Twp. 1 - 17, Rge. 15 - 30 W2M), northeastern Montana (Twp. 36N - 37N, Rge. 44E - 58E) and northwestern North Dakota (Twp. 162N - 164N, Rge. 102W - 103W). Geophysical well logs from 1,242 boreholes were examined (Fig. 2). Structure, isopach and second-order residual maps were utilized in the recognition and description of the complex chronologic history of salt dissolution in the area. Previous descriptions of different aspects of the study can be found in McTavish and Vigrass (1987) and McTavish (1990).

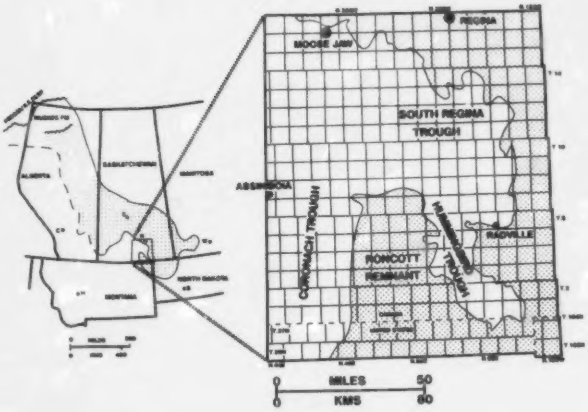


Figure 1: Location map showing regional and local distribution of Prairie Evaporite (After Holter, 1969; McTavish, 1990). Note that the present areal distribution of the Prairie Evaporite is patterned.

Figure 2: The area of investigation showing distribution of wells penetrating the Second White Speckled Shale.

GENERAL GEOLOGY

Phanerozoic strata of the northern Williston Basin can be divided into three gross lithologic units (Christopher *et al.*, 1971). The basal clastic unit comprises the sandstone and shale of the Deadwood and Winnipeg Formations (Cambrian to Middle Ordovician age; Fig. 3). The middle carbonate - evaporite unit embraces dolomite, limestone and evaporite assigned to the strata from the Red River Formation to Big Snowy Group (Middle Ordovician to Mississippian). The upper clastic unit includes sandstone and shale from the Watrous Formation to the glacial drift (Triassic or Jurassic to Quaternary age; Fig. 3). Several major unconformities are present resulting in significant spans of geologic time not being represented in the northern portion of the Williston Basin (Fig. 3).

The Williston Basin is the dominant structural feature of the area. Regional dips on the Precambrian approximate 6 m/km (32 ft/mile) to the south. Within the area of investigation preserved Phanerozoic rock thicknesses range from 2,000 m (6,500 ft) in the north to 4,000 m (13,000 ft) in the south. Numerous low relief basement structures of apparent diastrophic origin interrupt the simple regional structure. The 'Outlook structure' is an example of a basement dome of Devonian age (Wilson *et al.*, 1963). Most other local structures in post-Middle Devonian strata are associated with salt removal and collapse.

Evaporitic strata, mainly halite and anhydrite, occur at numerous stratigraphic levels in the Williston Basin. The Middle Devonian Prairie Evaporite is the thickest and laterally the most extensive of these layers (Fig. 3). The Prairie Evaporite reaches thicknesses of up to 100 m (300 ft) within the study area. Other Mississippian and Devonian salts which occur at numerous locales in the Williston Basin tend to be thinner and cover a smaller geographic area.

In some areas the Prairie Evaporite is absent due to post-depositional salt removal. Subsurface dissolution of 10s to 100s of metres of evaporite resulted in the collapse of the overlying rock column and the production of surface depressions which accommodated accelerated sedimentation. The process occurred in a number of phases resulting in structural complexities within the Phanerozoic fill of the Williston Basin. Many of the salt-removal structures are situated in a large area on the northern flank of the basin where dissolution of the Prairie Evaporite is complete (Fig. 1). Prominent salt-free locales include the Hummingbird Trough (Demille *et al.*, 1964), Coronach Trough (Broughton, 1977 and 1979; Whitaker *et al.*, 1978) and South Regina Trough. The Roncott Remnant (Roncott anticlinorium of Christopher, 1961) is a large north-ward projecting salient of preserved Prairie Evaporite (Fig. 1).

SUBSURFACE DISSOLUTION AND ITS RECOGNITION

Subsurface dissolution is the process by which evaporitic strata are dissolved by undersaturated fluids causing the removal of evaporites and a corresponding increase in the salinity of the formation waters. The removal of evaporites may create a void in the subsurface resulting in the fracturing and collapse of the overlying strata. In the process, various sized blocks collapse to fill the void. An example of the breccia produced during the process is seen to the northwest of the study area at Lsd. 13-4-25-8 W3M (Plate 12a of Holter, 1969). Slumping of the overlying beds produced lows at a paleo-surface in which compensating thicknesses of sediment may have accumulated.

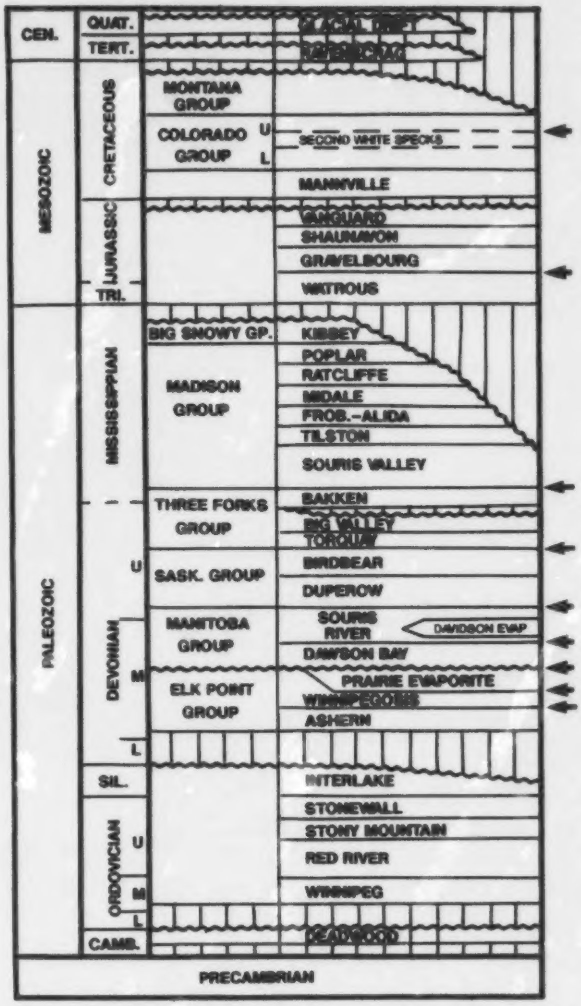
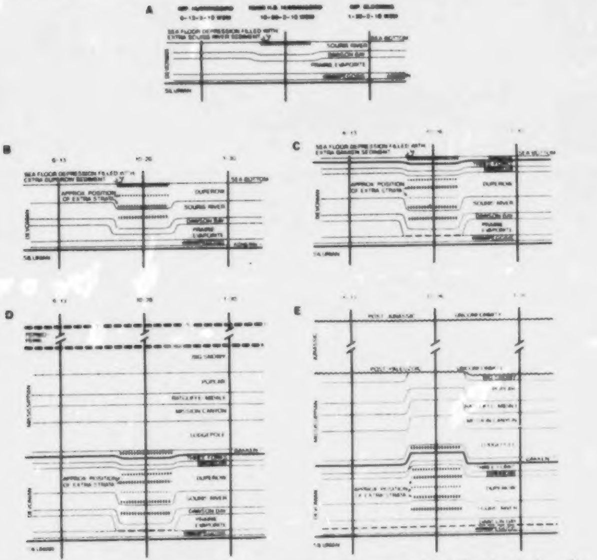


Figure 3: Rock-stratigraphic units in south-central Saskatchewan. Arrows show the horizons utilized in the construction of isopach and structure maps.

Salt dissolution features in the Williston Basin were first recognized by Baillie (1953) and Bishop (1954). Initially, the process of salt dissolution was believed to be one single event of evaporitic removal producing large local solution lows and increased thicknesses of sediment fill. Smith and Pullen (1967), and Swenson (1967) demonstrated that more complex localized domal structures were produced by multiple-stage salt solution and collapse. At the Hummingbird oil field (Twp. 2, Rge. 19 W2M), Smith and Pullen showed that the dome which forms a structural trap occurs as a result of local subsurface dissolution of the Prairie Evaporite with thickening of Devonian and Mississippian strata coupled with later but similar subsurface dissolution with thickening of post-Mississippian strata in the surrounding area (Fig. 4a-d). This later dissolution and collapse resulted in the original thickened strata being raised above the collapsed flanking beds (Fig. 4e).

Dissolution features originated as narrow local channels in the Prairie Evaporite perpendicular to the existing salt edge. The width of the channels increased with time until they coalesced to produce escarpments at the salt edge (Kent, 1968). Christopher (1961) suggested that dissolution channels shifted laterally as the collapse breccias were cemented, thereby blocking fluid flow in the original solution channels. The process was not necessarily continuous, but involved episodes of varying duration when little or no salt was removed (Kent, 1968). Some dissolution features in the Souris River salt (Davidson Member) coincide with areas of dissolution within the Prairie Evaporite (Lane, 1964).


A
B
C
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SEA FLOOR DEPRESSION FILLED WITH EXTRA SOURIS RIVER SEDIMENT
SEA FLOOR DEPRESSION FILLED WITH EXTRA DUPEROW SEDIMENT
SEA FLOOR DEPRESSION FILLED WITH EXTRA BAKKEN SEDIMENT
SEA BOTTOM
APPROX. POSITION OF EXTRA STRATA
DUPEROW
SOURIS RIVER
DAWSON BAY
PRAIRIE EVAPORITE
DEVONIAN
SILURIAN
BIG SNOWY
POPLAR
RATCLIFFE MEMBER
MISSION CANYON
LODGEPOLE
BAKKEN
MISSISSIPPIAN
JURASSIC
POST JURASSIC UNCONFORMITY
POST PALEOZOIC UNCONFORMITY

Figure 4: Schematic cross-section from the Hummingbird Field relating the development of multiple-stage salt dissolution structure during Devonian Souris River (a), Devonian Duperow (b), Mississippian Bakken (c), post-Bakken Paleozoic (d), and late Jurassic- early Cretaceous time (e) (Recreated after Figs. 8, 9, 10, 11 and 12 of Smith and Pullen, 1967).

EVOLUTION OF SALT DISSOLUTION FEATURES AND THE SALT-FREE AREA

In this investigation, dissolution events and their ages were based on subsurface mapping techniques including isopach, structure contour and residual maps. Maps were constructed using several conformable Phanerozoic lithic units with easily identifiable geophysical log character (Fig. 3). To minimize paleo-erosion effects on mapping, major unconformities were avoided as markers.

Isopach and structural anomalies can be sharpened by using residual maps derived from trend surface analysis. Second-order (second-degree) regional thickness and structure trends were removed producing second-order residual maps (Krumbein and Graybill, 1965). The mapping procedure and its use in the identification of the location and timing of salt removal and collapse is given by McTavish (1990).

Salt removal was initiated soon after deposition of the Prairie Evaporite and has continued to the Recent (post-Pleistocene). Figure 5 qualitatively illustrates the progressive growth of the salt-free area. Analysis of structure, isopach and residual maps suggests that the Prairie Evaporite was never deposited, at least not in appreciable amounts, west of the study area (106° W longitude).

At the onset of Madison Group deposition (Early Mississippian) the salt edge was very close to its original position with only minor areas of complete salt removal in the extreme northwest and along the western margin of the Roncott Remnant. Several salt thins were present throughout the Hummingbird and South Regina Troughs (Fig. 5a).

Between deposition of the Madison Group and Watrous Formation (Mississippian to Jurassic time) the entire Prairie salt was removed from large parts of the South Regina and southern Hummingbird Troughs (Fig. 5b). By the time of deposition of the Second White Speckled Shale (Late Cretaceous), Prairie Evaporite was entirely removed from the northern segment of the Hummingbird Trough (Fig. 5c). Much of this salt removal appears to be associated with the time represented by the prominent sub-Watrous (Jurassic) and sub-Mannville (Cretaceous) unconformities. The presence of such unconformities could pose a problem in distinguishing structure and isopach anomalies caused by salt dissolution from those due to erosion, compaction and the removal of minor Mississippian and Jurassic salts. McTavish and Vigrass (1987) noted that the Madison Group to Watrous Formation sediment package has seen minimal post-Middle Jurassic thickness modification from compaction and Madison salt removal. Thus, an isopach of this interval can be regarded as a paleo-structure map of the base of the Madison Formation during the Middle Jurassic. The anomalies within this interval are thus attributable to salt removal and collapse. Similar logic can be applied to the base of the Gravelbourg Formation to Second White Speckled Shale sediment package and the sub-Mannville unconformity.

Use of the subsurface mapping techniques may be demonstrated where salt removal has occurred since deposition of the Second White Speckled Shale. Latest Cretaceous and Cenozoic salt removal has taken place along a pronounced Second White Speckled Shale structural low. The negative anomaly extends south from Regina (Twp. 17, Rge. 19 W2M) to the Canada-United States border (Figs. 6a and b). The trough bifurcates at the northwestern end of the Roncott Remnant with one branch continuing south along the Coronach Trough and the other branch trending eastward to Twp. 6, Rge. 17 W2M. The east trending branch essentially divides the Hummingbird and South Regina Troughs. A small structural low (negative residual) occurs within a salt-free embayment in Twp. 14, Rge. 18 W2M. The negative residuals present in the southern Hummingbird Trough and area to the south and east are not dramatic and probably reflect a regional feature not attributable to salt dissolution.

EFFECT OF SALT REMOVAL ON OUTCROP DISTRIBUTION

The Missouri Coteau is an escarpment marking the boundary between the Second and Third Prairie Levels. The Second Prairie Level (Saskatchewan Plains) is underlain by Cretaceous shales and lies topographically lower than the Third Prairie Level (Alberta Plains), which is largely underlain by sandy Upper Cretaceous and Tertiary strata. Kupsch (1958) originally suggested that the Coteau is a fault-generated escarpment formed by initial faulting, differential compaction and later fault reversal (Fig. 10 of Kupsch, 1958). Many exploratory wells have been drilled since 1958 and the additional subsurface data suggest that the position of the Missouri Coteau is unrelated to underlying structure. The surface position of the Coteau (Fig. 6c) is not directly related to Second White Speckled Shale structure (Fig. 6a) but it appears that the position of an embayment within the Missouri Coteau at the southeastern end of the South Regina Trough (Twp. 6, Rge. 19 W2M) is influenced by post-Second White Speckled Shale solution collapse.

In somewhat greater detail, dissolution appears to have influenced outcrop distribution. Anomalous outliers and inliers of

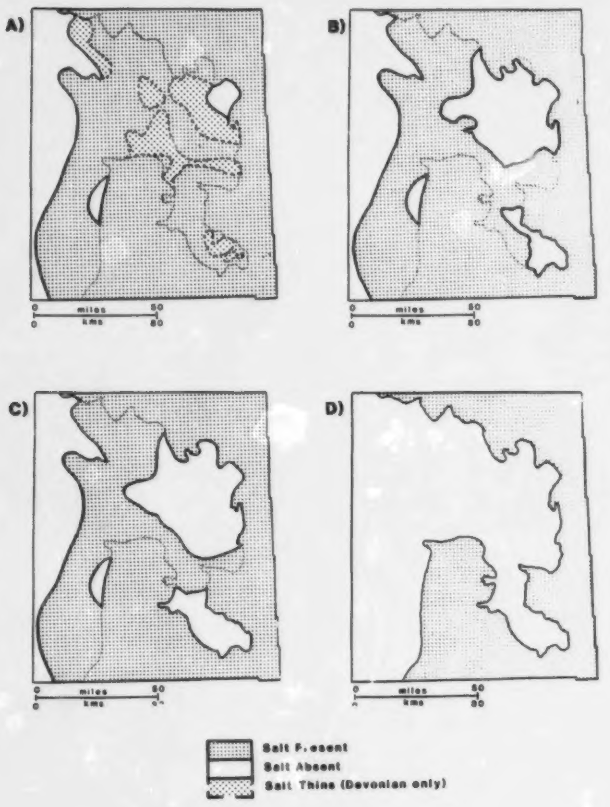

Figure 5: Development of the salt-free area. Maps show distribution at: A) Early Mississippian (at start of deposition of the Madison Group); B) Middle Jurassic (Watrous Formation); C) Late Cretaceous (Second White Speckled Shale); and D) Recent (present day).

strata are related to post-Second White Speckled Shale dissolution trends. The inliers and outliers were formed when structural collapse occurred during or after deposition of the sedimentary unit which now occupies the outlier or surrounds the inlier. Pleistocene glaciation and its associated erosional effects then caused a smoothing of the earth's surface. It is this combination of salt generated collapse and the erosional effect of glaciation which cause the conspicuous inliers and outliers in the area.

Post-Second White Speckled Shale dissolution and collapse trends (Fig. 6b) are interpreted based on maps such as the second-order residual structure on top of the Second White Speckled Shale (Fig. 6a). The dissolution trends are based on negative residual values corresponding to structural lows. Numerous inliers and outliers (Figs. 6c and 6d) occur within the area of interpreted post-Second White Speckled Shale dissolution.

Anomalous outliers of Ravenscrag Formation (Paleocene) occur north of the main belt of the formation along post-Late Cretaceous salt solution features (Fig. 6d). One prominent outlier occurs along the Coronach Trough in the Dirt Hills (Twps. 7 to 10, Rges. 22 to 27 W2M). A second outlier lies north of the Missouri Coteau near Radville (Twp. 6, Rge. 17 W2M). The second-order residual structure map of the Second White Speckled Shale (Fig. 6a) shows the outliers are in structural lows, indicating dissolution of the Prairie Evaporite post-dating deposition of the Paleocene strata.

Outliers of Wood Mountain Formation (Miocene) also occur in the Coronach Trough (Fig. 6d). Preservation of the Wood Mountain Formation in this area suggests that Coronach Trough dissolution was at least in part Miocene or younger.

Inliers of Bearpaw Formation occur southwest of the Missouri Coteau and the main belt of the formation (Fig. 6d). Most prominent of these inliers are the ones at Blackfoot Ridge on the Roncott Remnant (Twps. 5 to 7, Rges. 22 to 24 W2M) and west of the Coronach Trough (Twps. 7 to 9, Rges. 27 to 29 W2M). Second White Speckled Shale residual structure (Fig. 6a) shows both inliers have positive residuals and are structurally higher than the Coronach Trough.

Exact timing of post-Second White Speckled Shale salt removal is difficult to determine because good marker horizons are restricted areally. The examination of outcrop distribution within areas of known post-Second White Specks dissolution is one method which can improve the evaluation of the timing of solution collapse. Other lines of evidence which aid in the determination of the timing of post-Second White Specks dissolution have been previously recorded. It has been noted that the Ravenscrag Formation thickens in the Coronach Trough (Broughton, 1979; Whitaker *et al*., 1978) indicating that at least some salt removal and collapse occurred during the early Tertiary (Paleocene). Depressions on the eroded surface of Tertiary and Cretaceous bedrock, now infilled by Quaternary sediments, are common in southern Saskatchewan. Several of these depressions

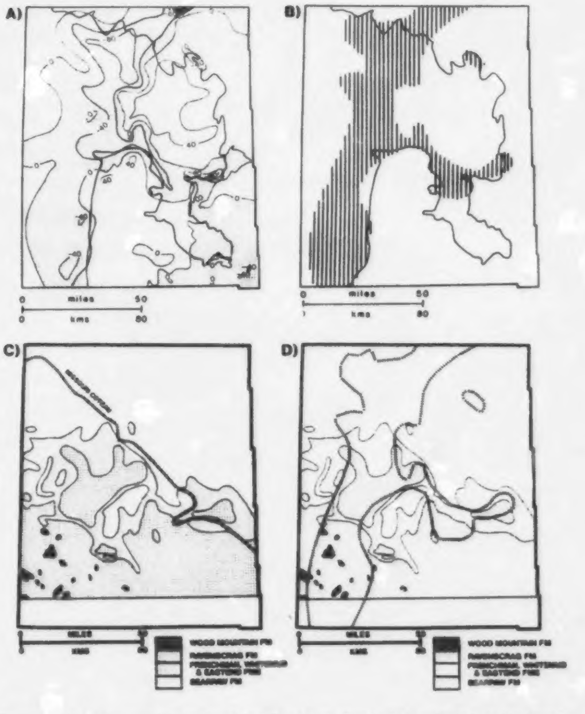

Figure 6: Post-Second White Speckled Shale salt dissolution and outcrop geology in south-central Saskatchewan; A) Second-order residual structure map on top of the Second White Speckled Shale. Contour interval is 40 m and negative residuals are patterned; B) Interpreted area of Prairie Salt solution and collapse that occurred after deposition of the Second White Speckled Shale (vertical ruling). Based on map 6A; C) Outcrop distribution within the Canadian part of the study area (based on MacDonald and Broughton, 1980); D) Area of post-Second White Specks salt dissolution, (dashed line) showing relationship to surface geology, inliers and outliers.

are related to subsurface dissolution of the Prairie Evaporite during the Pleistocene or Recent Epochs (Christiansen, 1967, 1971a, 1971b).

DISCUSSION: APPLICATION TO PETROLEUM EXPLORATION

Similar methodology can be applied to subsurface unconformity surfaces. At any major unconformity surface such as the sub-Jurassic and sub-Cretaceous, inliers or outliers have in the past been interpreted to be an erosional remnant. Inliers and outliers, however, can also potentially be formed by dissolution and collapse followed by erosion. Where petroleum bearing reservoirs subcrop at the unconformity surface these combined dissolution/erosion features could have a distinct trend and be of economic significance. Knowledge of subcrop trends coupled with detailed information on the timing of dissolution would be a useful addition to one's exploration strategy.

CONCLUSIONS

1) Salt dissolution and collapse is a major structure forming process in the southern Saskatchewan segment of the Williston Basin.
2) The multi-stage salt dissolution phenomenon has been active in the study area from the Middle Devonian, immediately after deposition of the Prairie Evaporite, to the Recent.
3) Distribution of outcrop in southern Saskatchewan has been influenced by Cenozoic salt dissolution and collapse coupled with the erosional effect of Pleistocene glaciation. This model may also be applicable to subcrop distribution at major unconformities such as the sub-Cretaceous and sub-Jurassic.
4) The examination of outcrop distribution may also aid in the interpretation of the timing of late-stage salt dissolution and collapse.

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THE RECOGNITION OF BASINAL RHOMBOIDAL FAULT PATTERNS AS A BASIS FOR THE DEVELOPMENT OF A PLATE TECTONICS STRUCTURAL MODEL FOR THE WESTERN CANADA SEDIMENTARY BASIN

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ABSTRACT

A new plate tectonics structural model is proposed for the development of the Western Canada Sedimentary Basin (WCSB). This model is based upon the recognition of complex regional rhomboidal fault patterns using Thematic Mapper imagery. Identification of detailed megafracture-fault (zone) patterns by applying selective DEMOFRAC methodology is the essential element in this modelling.

Recognizable patterns represent the regional system of fault zones which delimit the tectonic megablocks, thus forming a distinctive regional rhomboidal to off-orthogonal pattern which characterizes each particular fault or a zone. The complex regional system of faults and fault zones as revealed on the surface reflects the long structural history of the Basin. It has been controlled by the action and interaction of two dominant structural systems, repeatedly reactivated basement faulting and the "thin skinned" tectonics of the overlying sedimentary sequences, and of several tectonic episodes. These systems and episodes have, in turn, been controlled and driven by major plate tectonics events associated with the spreading of the Pacific Ocean floor during the Paleozoic and, later, with the opening and consequent spreading of the Atlantic and Arctic Ocean floors during the Mesozoic and Cenozoic.

Westward transpressional underthrusting of the Precambrian Shield Plate (PSP) beneath the Cordilleran Semi-Plate (CSP) is the basic mechanism of the proposed structural model for the WCSB . This underthrusting resulted from the westward continental drift and collision(?), and was combined with rotation of the PSP relative to the CSP. Subhorizontal detachment zones have developed within the shale and/or other ductile formations beneath "rigid packages" of thick sedimentary sequences during successive episodes of underthrusting, as a result of interaction of recurrent compression with "thin skinned" tectonics. As a consequence, an apparent reverse sliding (or "sliding back") of the overlying sequences seems to have occurred. These sequences would "lag behind" the underlying sequences due to hindrance from the barrier - The Rocky Mountains Thrust and Fold Belts. The final result is a complex and partly displaced surface regional fault pattern as compared with the basement fault pattern.

Due to the existence of local stress fields compressional and extensional structures and features occurred simultaneously with the rotation and lateral shifting of tectonic megablocks. Such conditions should have been more intensive closer to, and along the edge of the "Disturbed Belt".

INTRODUCTION

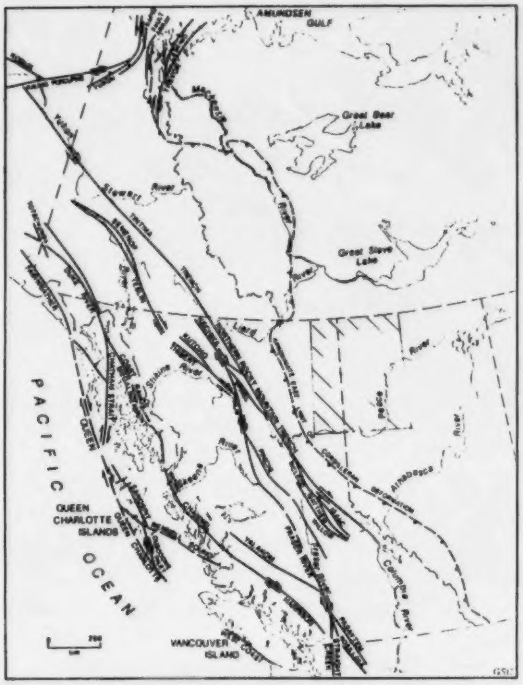
During his work in structural geology and photointerpretation for various exploration projects, the author has concluded that the "single straight line" symbol for characterization of faults is apparently an oversimplification and insufficient as a exploration tool. According to his findings, this symbol should be modified and/or diversified to reveal the actual details of each particular fault. Although segments of faults (zones) may often be straight for some distance, it is the megafracture-fault pattern that determines their nature and character. Semiquantitative interpretation and idealization of the "fault trace" by "lineament analysis" cannot satisfy the requirements of the explorationist in his search for the subsurface structures which are capable of trapping hydrocarbons.

To fulfill such a need, the author has used a combination of common symbols for the fault (zone) pattern. Thin lines denote secondary conjugate faults and major megafracture patterns, thick lines indicate the major conjugate faults and main faults within the zone, while shading envelops the extent of a particular fault zone (Fig. 2).

THE STUDY AREA AND THE BASIS FOR INTERPRETATION

The north-central parts of the WCSB within the regions of northwestern Alberta and northeastern British Columbia (116° to 122° West and 56° to 60° North) have been chosen as the area for study (Fig. 1). There are several well explored major oil and gas fields (such as Zama and Rainbow) within the region and these are surrounded by huge areas with much less well and seismic controls. Judging the prospectiveness of these under-explored areas from the standpoint of the newly developed structural model, it is possible to predict with some certainty the existence of undiscovered hydrocarbon fields, particularly the ones with dominant tectonic control.

The basis used for the collection of data has been an excellent quality and higher resolution satellite imagery from the Landsat 5-TM (Thematic Mapper). A colour composite of bands 3, 4 and 5 (Blue, Red, Green) has been used, because it proved (besides band 7) to be the most useful for detection of megafracture-fault patterns and anomalous features. These linear to curvilinear features are essential for identification of the regionally predominant rhomboidal fault zone (patterns). These patterns bound and structurally control the tectonic megablock system within the Plains region. In order to match the regional topographic, geological and other maps, prints of TM-imagery were enlarged to the scale of 1:250,000.



Distribution of the main transcurrent faults in the Canadian Cordillera and adjacent Alaska

Figure 1: Location map (adapted from Gabrielse and Yorath, 1989).

METHODOLOGY USED

The author has developed and successfully applied the DEMOFRAC method to several exploration projects in the search for hydrocarbons and minerals, both in Canada and internationally. Several types of analyses are employed: geomorphological, drainage pattern, megafracture-fault pattern, recent erosion, uplift and redepositional processes, vegetational and tonal anomalies in combination with some less apparent types. A widely accepted misconception that glaciation processes in the Plains regions mask structural information on the surface is rectified.

DEMOFRAC follows three successive but distinct stages: 1) analytical selection of specific data; 2) placing data into a regional geological framework, development of structural models and classification of anomalies; and 3) synthesis. Application of this method reduces the need for shooting seismic grids by concentrating exploration on the most prospective localities. It also results in considerable savings of time, effort and money. Significantly DEMOFRAC detects megafracture-fault linear to curvilinear features, the length of which may range from a few to many scores of kilometers. In this respect the DEMOFRAC approach differs from lineament and other kinds of fracture analysis.

GEOLOGICAL OVERVIEW

The existence of "global regmatic fractures" has been postulated for some considerable time and has been accepted by numerous professionals. Among others, Thomas (1974) mapped the pattern of "regmatic lineament-fractures" in northern Montana, and measured their intersecting angles. Greggs & Greggs (1989), and Thomas (1974) postulated the importance of Precambrian basement controls on the "rhegmatic fracture pattern". This author's interpretation, using thematic mapper imagery, revealed much more complex and somewhat intriguing regional rhomboidal fault (zone) patterns on the surface, and broadened the concept of the above-mentioned "basement controls" to stress their interaction with "thin skinned" tectonics.

On the other hand, most publications on the Western Canada Sedimentary Basin follow the established opinion about deformation within this basin. The popular yet apparently misleading "Approximate East Limit of Cordilleran Deformation" or "Eastern Edge of the Disturbed Belt" implies that regions west of this line are "disturbed" (Thrust and Fold Belts), as opposed to the "undisturbed" or locally disturbed region to the east, which is the major part of the basin. The exceptions within this eastern region are in the Peace River Arch area and around some oil and gas fields, where normal faulting has been detected by geophysical and other methods.

While numerous photogeological and lineament analyses, including this study, provide indications that the Plains region of the WCSB is considerably disturbed, but by different mechanisms and styles, the tectonic and other geological maps do not display this in general.

REGIONAL RHOMBOIDAL FAULT (ZONE) PATTERNS AND THE SYSTEM OF TECTONIC MEGABLOCKS

There are few professionals, and even fewer publications, that recognize the dominant role of rhomboidal fault (zone) patterns in the development of the WCSB. The general trends of these fault zones are NE-SW, with some verging to the north and NW-SE with vergence to the west. Thomas (1974) concluded that: "oil and gas localizations are closely associated with drag folds or lineaments", and "ore localities suggest similar associations" to his tectonic "block framework". Thomas's conclusions are the closest to the findings revealed by the DEMOFRAC method which are explained in this paper. A similar, rhomboidal to triangular pattern of "linears" from Alaska was published by Lathram and Albert (1974). In their study of Paleozoic sedimentation, Sikabonyi and Rodgers (1959) concluded that the predominant NE-SW trending basement faults controlled the regional diversification of sediments and their facies. A fault controlling mechanism for some oil and gas fields has been postulated by some geologists (*e.g.*, Jones, 1980). Greggs & Greggs (1989) published a model of fault block controls for the development of Devonian reefs. The major northeast-trending fault zone which cuts the study area diagonally was identified geophysically and named by Hoffman (1987) "The Great Slave Lake Shear Zone." Parts of this zone to the southwest are also known as the Hay River Fault.

The regionally dominant rhomboidal pattern of fault zones dissects the Plains region of the WCSB into a system of tectonic megablocks. The average size of these blocks is approximately 160 by 200 (± 40) km. The width of fault zones varies greatly from a few kilometers to over 20 km across (there being one exception - close to 70 km). The peripheries of major tectonic blocks have been cut by conjugate sets of secondary faults into a mosaic of minor tectonic blocks. Their shape varies from rhomboidal to trapezoidal and triangular. The last are developed close to the intersections of major fault zones. Their average size is 20 by 30 (± 5) km.

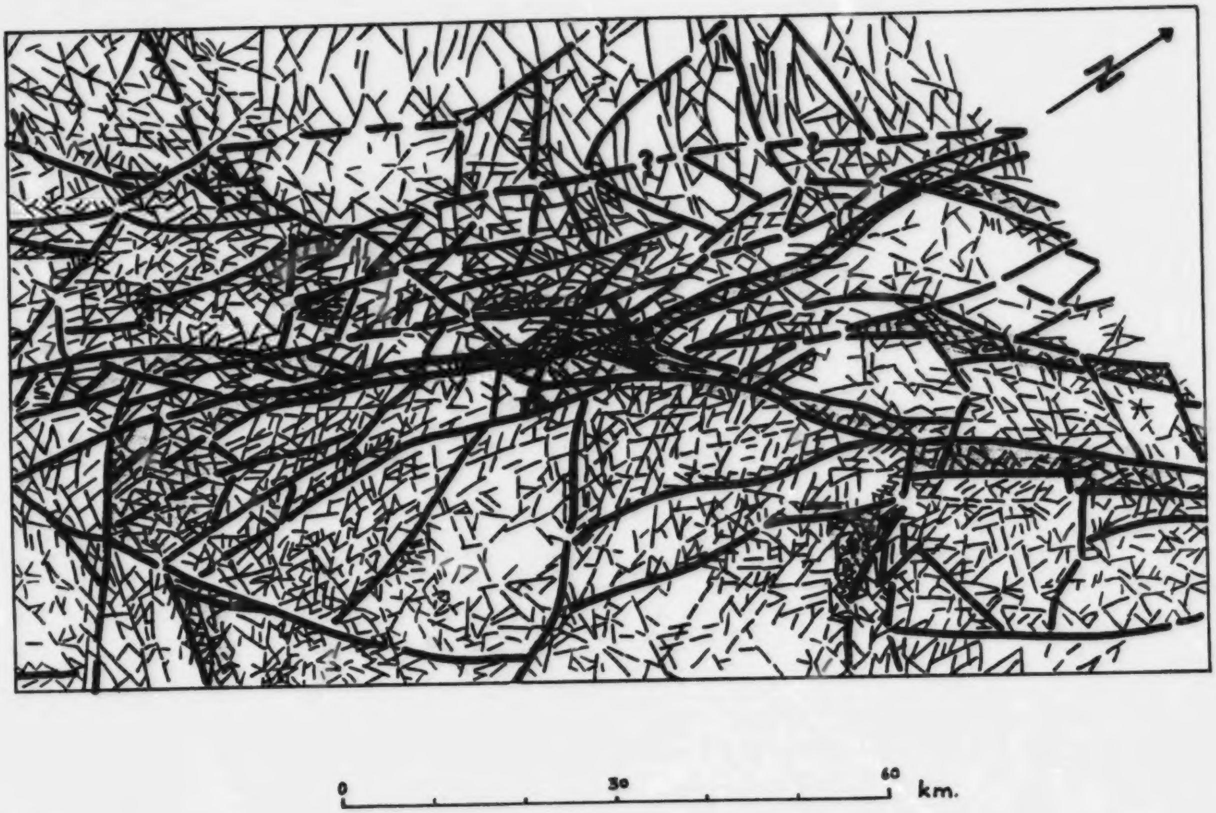


Figure 2: Megafracture-fault patterns in northern Alberta.

The significant prevalence of horizontal (wrench type) displacements along most fault zones has been observed through the analysis of fault patterns by the combination of ellipsoids of deformation.

The regional pattern of repeatedly reactivated basement faults has played a major role throughout the geological history of the WCSB, and in particular has controlled: diversification of sediments and their facies; structural development of reservoirs for entrapment of hydrocarbons (with associated other types); development of secondary-fracture porosity; migration of hydrocarbons along concentrated pathways; and secondary migration and consequent depletion of some reservoirs.

POTENTIAL SUBHORIZONTAL DETACHMENT (DECOLLEMENT) ZONES IN THE SUBSURFACE OF THE WCSB-PLAINS REGION

There is a potential for development of more than one subhorizontal detachment zone below the rigid packages of thick sedimentary sequences, *i.e.* along zones of incompetent-ductile underlying sediments. There are at least four major (out of a maximal number of eight) such rigid and competent, thick sedimentary "packages". These have been deposited over the Precambrian Shield in a "multi-layered cake" fashion, and over other incompetent sediments. Regional detachments have most probably evolved along these latter beds due to their lithology, directional compression and high fluid pressures. A careful study of the Stratigraphic Correlation Chart and the Lexicons of Canadian Stratigraphy for Western Canada strongly indicates where these detachment zones could have developed (see the arrow indicators in Fig. 5). Six such potential zones have been identified. All of them may not have been developed within the entire Plains region, but may be within some tectonic megablocks. Besides this limitation, only a few, or maybe one zone may have been active during a certain geological time span. It logically follows that lower zones would have been active earlier than the overlying ones.

These potential subhorizontal detachment zones, as viewed from the uppermost and youngest, are as follows:

ZONE I:	UPPER CRETACEOUS - shales along the upper parts of the Smoky Group (Colorado Group), Paskwaskau, Lea Park or Pakowki shales;
ZONE II:	LOWER JURASSIC - shales within the Fernie Group - Black Shale or Poker Chip shale, and UPPER JURASSIC - Passage Beds shale;
ZONE III:	UPPER DEVONIAN to LOWER CARBONIFEROUS - shales within Exshaw Formation, top of Fort Simpson or Besa Formation (to the S.W.);

ZONE IV: UPPER DEVONIAN - shales within the top of Ireton or Muskwa Formations (Woodbend Group to the S.E.);

ZONE V: It includes possibly three formations - MIDDLE DEVONIAN Watt Mountain Formation, Cold Lake Salt and LOWER DEVONIAN - Basal Red Beds (confined to Central Alberta);

ZONE VI: MIDDLE CAMBRIAN - shales within the Snake Indian Formation - Stephen Member.

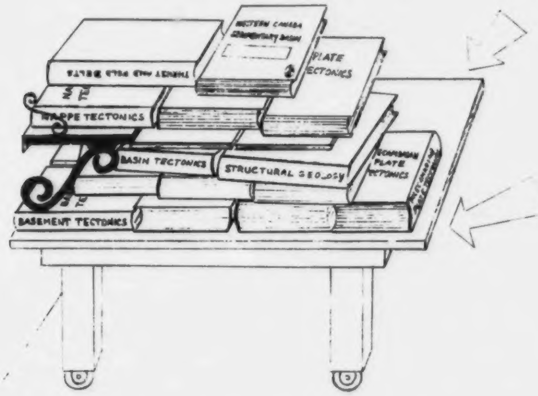


Figure 3: Test-sketch for the structural model of the Western Canada Sedimentary Basin.

PLATE TECTONICS STRUCTURAL MODEL OF THE WCSB

The author demonstrates his structural model by using a moving table covered with books (instead of applying the low angle snow plow effect, as used for years before) (Fig. 3). This is due to the similarities between results of this study about the structural build-up for the WCSB, with books piled in layers on a table (in "multilayered cake" fashion). The vertical piles of books would represent tectonic megablocks with vertical spaces between them (fault zones). Each horizontal layer of books would represent a rigid "package" of thick sedimentary sequences, and the horizontal space between them would depict subhorizontal detachment shale zones. The table is the Precambrian Shield Plate (PSP) moving westward (underthrusting), and the books (tectonic megablocks) are restrained from moving with the table (PSP) by resistance from the shelf barrier on the wall (Rocky Mountains and Foothills Thrust and Fold Belts). As a consequence, the upper layers of books (rigid sedimentary sequences) would "lag behind" the movement of the lower layers of books and the table (underlying rigid sedimentary sequences and the Precambrian Shield).

With change of table movement (compression direction) and push from the side (transpression), some layers of books (rigid sedimentary sequences) would slide laterally and their parts would rotate relative to one another (individual tectonic megablocks due to local stress field).

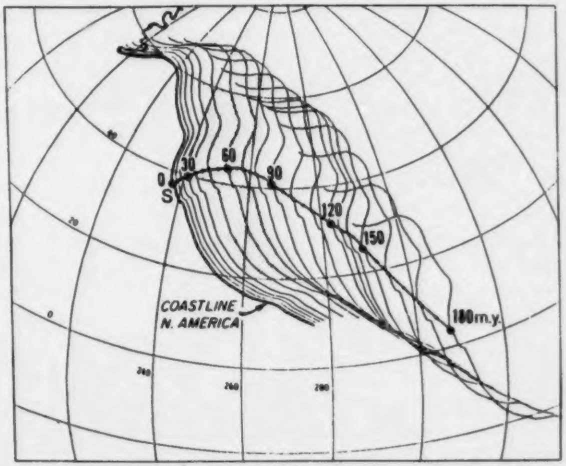
The author has applied a similar line of thought and a mechanism from the above test-sketch to structural modelling within the WCSB. This has resulted in the following model and description of its dynamics.

The basic mechanism of the proposed Structural Model for the WCSB is explained as the general westward transpressional underthrusting of the Precambrian Shield Plate (PSP) beneath the Cordilleran Semi-Plate (CSP). The transpressional components of underthrusting are explained by the northwestward left-lateral movement of the PSP (relative to the CSP), combined with a clockwise rotation since Lower Jurassic times. Later, during the Cenozoic, subsequent southeastward right-lateral movements occurred in combination with the anticlockwise rotation of the PSP (relative to the CSP). At present the North American continent is moving 2 to 2.6 cm/yr in a southwesterly direction, and is probably rotating in a clockwise manner again.

The above-described general westward movements of the North American continent have been caused and closely controlled by the opening and subsequent spreading of the Atlantic Ocean floor. The "motion of North America with respect to the hotspots for the last 180 M.a." has been ingeniously illustrated by Engebretson (1982) (Fig. 4).

The major transcurrent movement between the PSP and the CSP measures many hundreds of kilometres. This most probably has occurred along the Rocky Mountains Fault System (RMFS) within and/or east of the Omineca Belt. The fault system is composed of arrays of fault zones, including those of the Tintina and Rocky Mountain Trench. The reactivation of movements along a given fault or a fault zone would be the result of changes in the regional stress field's direction during various tectonic episodes. The slip along faults may occur in a direction different from that which took place in its earlier geological history (as previously mentioned).

Due to the existence of local stress fields along and between tectonic megablocks, the development of compressional and extensional structures should have occurred simultaneously. Such situations should be more pronounced closer to, and along the edge of the "Disturbed Belt".



Motion of North America with respect to the hotspots for the past 180 Ma in 10-m.y. intervals. The dark line with closed circles is the trajectory of a point riding with North America near present-day San Francisco (S) at 5 m.y. intervals with the larger closed circles at 30 m.y. intervals. Shaded region indicates time span of Laramide orogeny.

Figure 4: Curve of motion for North America during last 180 M.a. (from Engebretson, 1982).

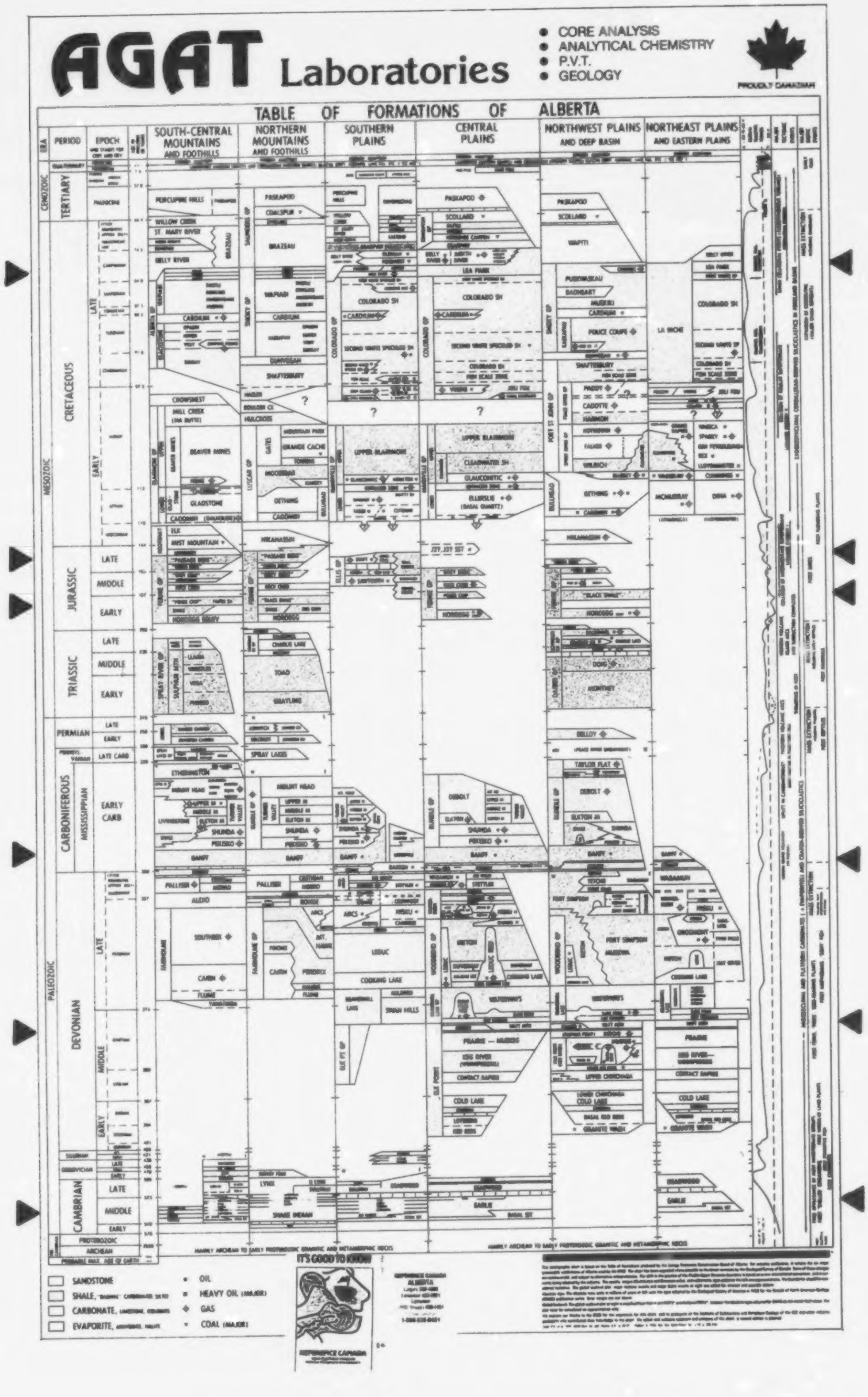


Figure 5: Proposed detachment zones in the Western Canada Sedimentary Basin.

As the PSP "approached"(?) and was "colliding"(?) with the barrier (Cordilleran Semi-Plate), the development of subhorizontal detachment zones occurred within the shale formations. Such detachment zones may be found below the major thick, rigid sedimentary sequences (Fig. 5). With the continuation of underthrusting of the PSP, the top sedimentary sequences would "lag behind" and thus "slide back" relative to the underlying sequences along these detachment zones. The transpressional-lateral movements of the PSP would also induce and/or cause lateral shifting and rotation of tectonic megablocks and their rigid sequences relative to one another.

The final consequence of transpressional underthrusting of the PSP was a development of much more complex and widespread patterns of faults and fault zones within the overlying sedimentary sequences. These have been propagated to the surface by repeated reactivation. Such patterns have been accentuated by glacial ice overloading, and by postglacial rebound. Due to the nature of glacial deposits, which behave like brittle rocks, most of these patterns and their detailed systems have been propagated through to the surface. The Holocene and the most recent geological processes have accentuated these patterns which are readily revealed on satellite imagery.

The proposed structural model also explains why there is a poor correlation (or none at all) between some fault patterns and features at basement level and within the lower and upper sedimentary sequences in relation to the surface patterns. As an example, the original shape and position of Rainbow Basin would fit surprisingly well and much better into the surface fault pattern, being located several tens of kilometres to the northeast from its present position. This apparent discrepancy can now be logically explained by using the complex surface fault and anomaly patterns as the basis for interpreting the dynamics of sequential movements within the WCSB.

Regional rhomboidal fault patterns have also played a major role during the initial stages of ramps for most thrust faults within the disturbed belt. Furthermore, the simultaneous and contemporaraneous development of lateral transverse or cross faults, with associated features within transitional zones, have been controlled by the major conjugate faults within the same regional system of fault zones. Thrust fault ramps and transverse-cross faults would have been developed within these above-mentioned inherited fault patterns, because these zones had already been disturbed and therefore were the weakest tracts within the regional stress field.

The development of the "Triangle Zone" (double wedge-shape with opposing points) along the "eastern limit" of the "Disturbed Belt" has also played a significant role in the structural build-up of the basin. It controlled the uplift along the opposing-reverse and thrust faults above it, as well as the development of the "blind fault(s)" underneath. The apparent "sliding to the east" of some rigid packages of thick sedimentary sequences would have occurred in combination with this compressional faulting.

It might be sufficient for an experienced professional to examine a detailed relief model of Western Canada and recognize these transverse-cross faults as distinctive, short cross-valleys. By examining the Geological Relief Map of Alberta, the regional rhomboidal fault patterns and tectonic megablock system being developed to the east and away from the Rocky Mountains and Foothills regions can be detected (Fig. 6).

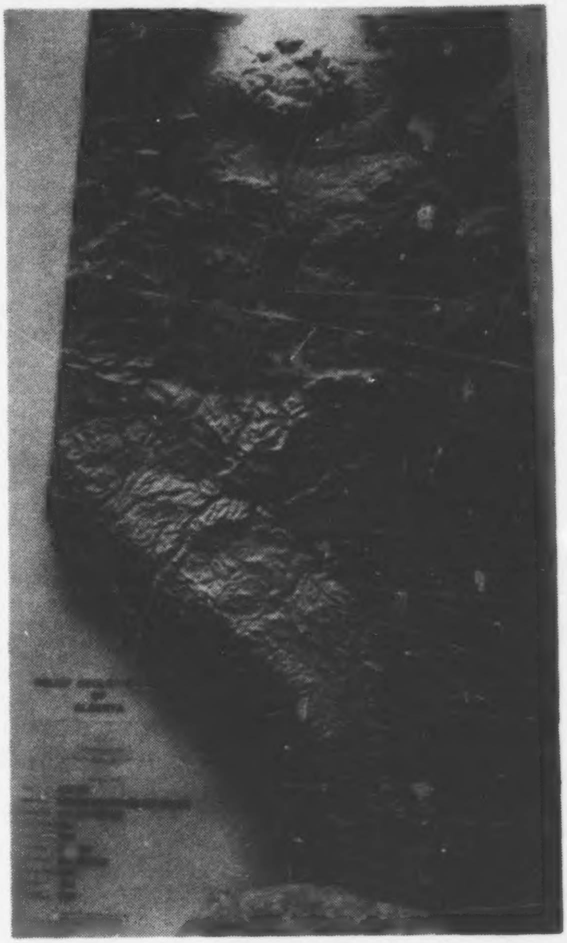


Figure 6: Relief geologic map of Alberta (Yager and Pascual, 1975).

CONCLUSIONS

The recognition of the importance of regional rhomboidal fault patterns, which delimit tectonic megablock structures, gives an exciting and promising new dimension to oil and gas exploration within the WCSB.

The concept of tectonic controls on migration of hydrocarbons and development of reservoirs, together with structural modelling has already assisted in the finding of new oil and gas prospects.

Although the WCSB has been very well explored stratigraphically, the proposed plate tectonics model essentially clarifies the less understood structural development of this mature basin. The prevalence of wrench type of displacement along fault zones and the importance of local stress re-adjustments between tectonic megablocks help to direct the modelling for hydrocarbon exploration towards structural types of prospects.

The well developed, complex and regional pattern of fault zones and their associated systems of megafracture-fault patterns indicate that tectonic controls may have had much greater influence over the migration of hydrocarbons along confined pathways. Tectonism has significantly influenced the regional and local deposition of sediments and played a dominant role in the development of structures, reservoirs and porosities within the basin.

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IMPORTANT NOTE:

Due to the confidential nature of the data and results which have been obtained during this study, in order to corroborate the ideas and confirm the conclusions reached, the author had to use illustrations from published sources.

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ANOXIC SULPHIDIC DIAGENESIS IN THE ORDOVICIAN WINNIPEG FORMATION OF SASKATCHEWAN

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ABSTRACT

Sandstones and shales of the Ordovician Winnipeg Formation from the subsurface of southern Saskatchewan contain remarkable textural and organic chemical evidence of anoxic sulphidic diagenesis. Several different types of sulphide-coated grains occur in the sandstones and are interpreted as having formed a few to several tens of centimetres below the water/sediment interface, near the base of the zone of bioturbation. Sulphide precipitation was brought about by bacterial reduction of sea-water sulphate to H_2S. Iron was present in the sediment as detrital oxide grains and probably also as bacteriologically-bound colloidal species. Work in progress shows that the oöids of the Winnipeg Formation contain high amounts of aminoacids, possibly representing partly degraded remnants of the organic matter which mediated the precipitation of the sulphides. Evidence of sulphidic diagenesis in the shales include: 1) abundance of framboidal pyrite; 2) formation of pyrite interstitially to Fe-oxide oöids which formed during oxic sedimentation; 3) replacement of microscopic algae with perfect preservation of chains of cells; 4) pyrite filling of chitinozoan tests; and 5) pyrite replacement of various fossil fragments.

Sulphidization textures that can be considered the precursors of those of the Winnipeg Formation occur in recent sediments of the Venice Lagoon, Italy, and of Long Island Sound. Anoxic sulphidic diagenetic environments must have been widespread in the shallow epeirogenic seas of the early Paleozoic of Canada since pyrite oöids are also found in the Great Bear Lake region and in Newfoundland.

INTRODUCTION

The Winnipeg Formation of Saskatchewan contains a variety of early diagenetic pyrite textures. In this paper the textures are discussed in the context of the diagenetic conditions in which they formed. This study is based on the examination of 50 polished-thin and polished sections, mostly from the middle part of the formation.

The most characteristic pyrite grains of the Winnipeg Formation are sulphide-coated grains (oöids and pisoids) which, while constituting somewhat of a mineralogical oddity, seem to be unusually common in early Paleozoic siliciclastic rocks of Canada (Fig. 1). Baillie (1952) reported the occurrence of pyrite oöids from outcrops of the Winnipeg Formation on Lake Winnipeg. Paterson (1971) recorded their presence in at least 35 petroleum exploration drillholes in southern Saskatchewan where the Winnipeg Formation was cored. Binda and Simpson (1989) described eight types of sulphide-coated grains from clay-rich sandstone of the formation in Saskatchewan. Work on the amino acid content of these oöids (Binda *et al.*, 1991) and on the comparison of sulphide textures of the Winnipeg Formation with those found in modern sediments of the Venice Lagoon, Italy (Frizzo and Binda, 1989) is in progress. Preliminary data from this research in progress are discussed here.

Other notable early Paleozoic occurrences of sulphide-coated grains of Canada include the Ordovician Wabana Formation of Newfoundland (Hayes, 1915; Kalliokoski, 1966) and a Cambrian (or Ordovician ?) occurrence in the Leith Peninsula (Great Bear Lake, N.W.T.) which is presently being studied by the author and by a graduate student.

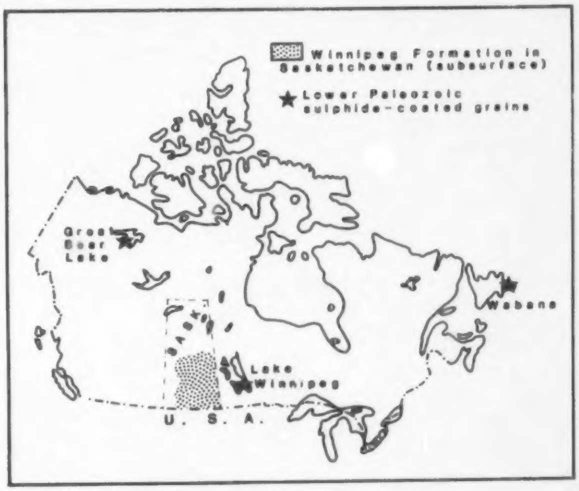
▨ Winnipeg Formation in Saskatchewan (subsurface)
★ Lower Paleozoic sulphide-coated grains
Great Bear Lake
Wabana
Lake Winnipeg
U. S. A.

Figure 1: Map showing Lower Paleozoic occurrences of sulphide-coated grains in Canada and areal extent of the Winnipeg Formation in Saskatchewan.

GEOLOGY

The Winnipeg Formation of Manitoba, Saskatchewan, and the Dakotas is a siliciclastic unit that reaches maximum thickness slightly in excess of 100 m in North Dakota and onlaps the Canadian Shield at about 54°30' Lat. N. Most of the formation is concealed beneath younger rocks. It is exposed on the islands and southern shore of Lake Winnipeg (Baillie, 1952) and, to the north, at the truncated onlap, in the Amisk Lake - Deschambault Lake region of Saskatchewan (Kupsch, 1952; Genik, 1954). The Winnipeg Formation marks the onset of Tippecanoe Sequence (Sloss, 1963) sedimentation in the ancestral Williston Basin.

The lower contact is either an angular unconformity on Precambrian schist and gneiss, or a disconformity on the

Cambrian-Ordovician Deadwood Formation. The Winnipeg can be differentiated from the Deadwood on the basis of higher textural and mineralogical maturity (Genik, 1954), lack of glauconite, and a lower radioactivity count on the gamma-ray log (Paterson, 1971). The upper contact with the Red River Formation may be gradational.

The age of the Winnipeg Formation is generally given as either Middle or Late Ordovician owing to the dearth of well-preserved macrofossils (Paterson, 1971). Work in progress suggests that a more precise definition may be achieved using acritarchs, chitinozoans, and scolecodonts which are fairly abundant in some shaly beds.

Paterson (1971) reviewed previous stratigraphic subdivisions of the formation and subdivided it into the Black Island Member, composed predominantly of quartzose sandstone with minor shale, and the overlying Icebox Member composed mainly of shale with minor, discontinuous arenaceous beds (Fig. 2). Vigrass (1971) by means of bed-to-bed correlation of drillholes in Saskatchewan and Manitoba, showed the existence of a predominantly sandy basin-margin facies in the north and of a basinal mudstone facies in the south. In Manitoba a sandy offshore bar facies is present in the upper part of the formation.

Sandstone of the Black Island Member is of two main types: 1) clean quartz arenite, either cemented by silica mobilized by pressure solution and precipitated in the pressure shadows, or poorly consolidated and friable; 2) quartz wacke containing 20-25 percent clay matrix. The quartz wacke is predominant in the upper part of the member. It is more pyritic and commonly bioturbated (Paterson, 1971). The sand fraction of the Black Island Member is composed largely of very well rounded, monocrystalline medium to coarse grains, probably recycled from the underlying Deadwood Formation. Angular to sub-rounded monocrystalline and polycrystalline quartz, together with rare feldspar grains suggest that the Precambrian Shield was also a source of detritus.

The Icebox Member is mainly a greenish waxy shale containing fragmental fossil material and commonly burrowed and mottled (Paterson (1971). At places it contains discrete layers of goethite oöids. A variety of pyrite textures occur in the Icebox Member.

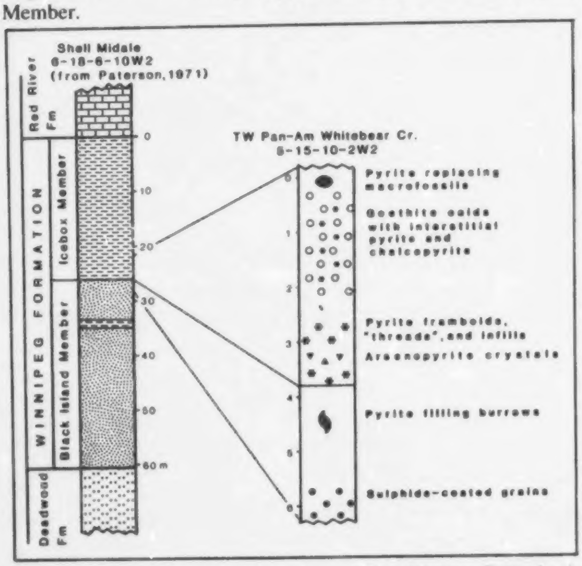

Figure 2: Generalized stratigraphic column of the Winnipeg Formation in Saskatchewan and occurrence of sulphide textures near the contact between the Black Island and the Icebox Members.

SULPHIDE TEXTURES

The occurrence of typical sulphide textures at the Black Island/Icebox Members boundary is shown in Figure 2. Some of them are illustrated in Figures 4 and 5.

FRAMBOIDS

Polycrystalline pyrite spheres (framboids) are abundant in the shale of the Icebox Member, but occur also in the clay matrix of wacke of the Black Island Member. Their average diameter in thin section is in the region of 7.5-10 µm. Their size distribution is unimodal positively skewed (Fig. 3). Only a few grains have diameter between 25 and 35 µm and they probably represent the tail of the skewed distribution rather than a separate mode. The diameter of the microcrysts that make up the architecture of the framboids ranges up to 2.5 µm; however, the large majority of framboids are composed of microcrysts with diameter less than 1 µm. Applying formulae given by Rickard (1970), it can be shown that individual framboids of the Winnipeg Formation contain between a few hundred and tens of thousand microcrysts. Frequently the microcrysts seem to be recrystallized and the cross-sectional area of the framboid appears as a massive pyrite surface, with only a few recognizable microcrysts (Fig. 4D). Recrystallization of framboids can occur early in the history of a sediment as it has been observed in recent pyrite framboids of the Venice Lagoon (P. Frizzo, *pers. comm.*). Framboids occur either isolated or in clusters and chains (Fig. 5F), and may become incorporated within sulphide-coated grains (see below).

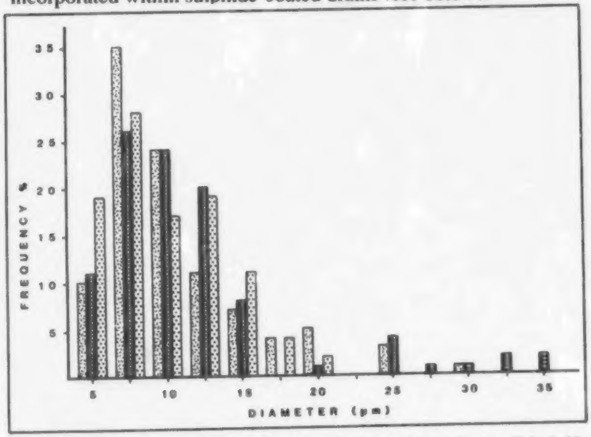

Figure 3: Size frequency distribution of pyrite framboids in three polished thin sections of shale from the Icebox Member (100 framboids per sample).

SULPHIDE-COATED GRAINS

Sulphide-coated grains may be found throughout the Black Island Member, but are abundant only in the more argillaceous wacke. Locally, particularly near the boundary with the Icebox Member, they are the major constituent of beds several centimeters thick (sulphidites). Eight distinct types of coated grains have been recognized by Binda and Simpson (1989), on the basis of accretion of a pyrite cortex around the following different types of nucleus:

(1) marcasite rosette;

(2) detrital quartz grain partially replaced by pyrite;

(3) skeletal fragment more or less replaced by cryptocrystalline pyrite;

(4) detrital ilmenite grain which may show sulphidization of hematite lamellae;

(5) chalcopyrite grain which may be of detrital origin;

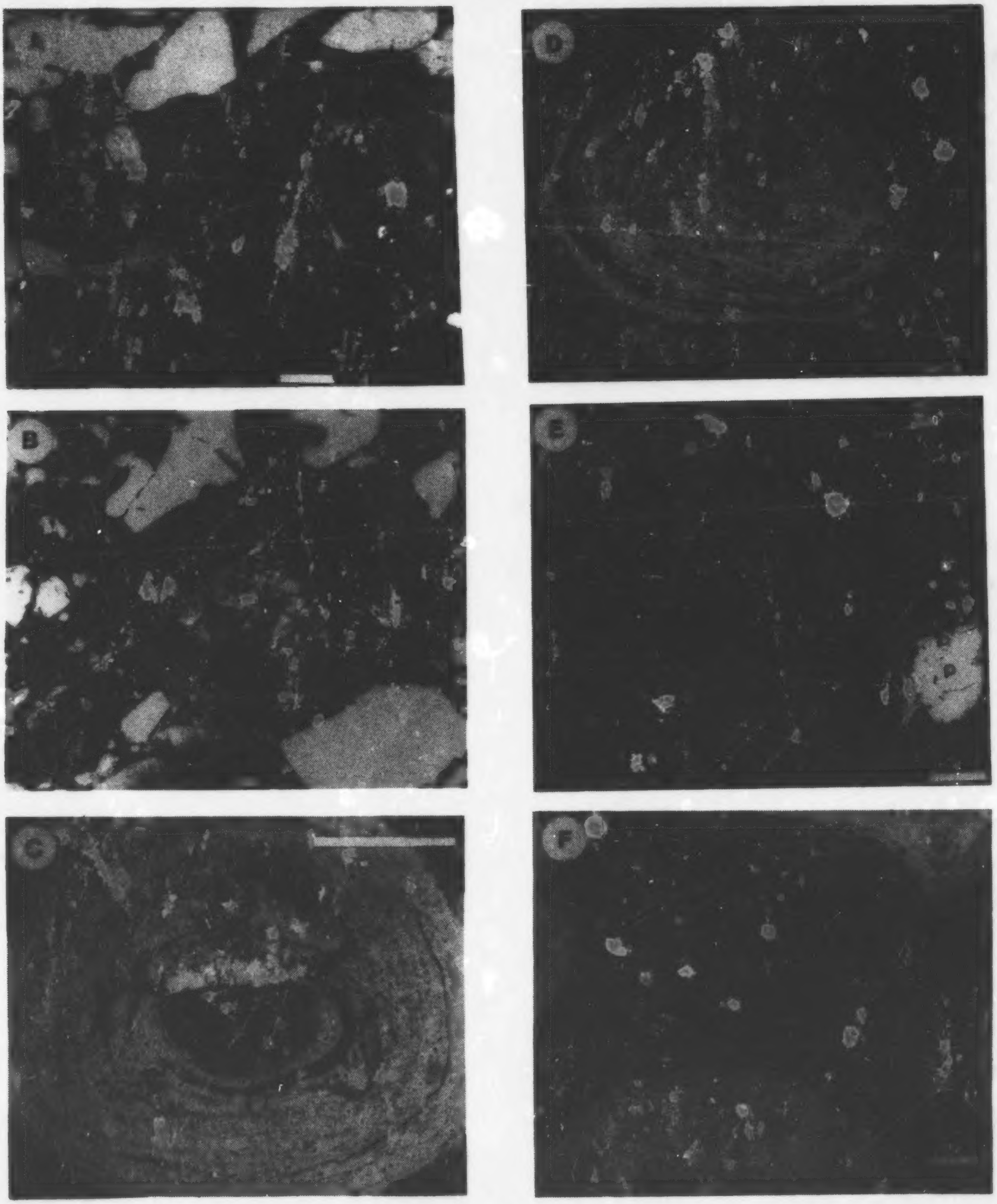

Figure 4: Photomicrographs of sulphide-coated grains (A-D) and of oölitic ironstone (E-F) in polished thin section of the Winnipeg Formation. A, B, E, F were taken with combined transmitted and reflected light, C and D in reflected light. Scale bar 0.1mm.

A: marcasite radiating aggregate (Black Island M.).

B: oöid with radiating marcasite (m) nucleus surrounded by pyrite (p) cortex (Black Island M.).

C: pyrite oöid with detrital ilmenite nucleus (Black Island M.). Tiny specks of pyrite oriented parallel to the long intercept of the grain formed by sulphidization of hematite lamellae.

D: pyrite oöid showing incorporation of framboids within the cortex (Black Island M.).

E and F: oölitic ironstone (Icebox M.). The oöids are composed of goethite (g). Pyrite (p) and chalcopyrite (c) occur in the matrix.

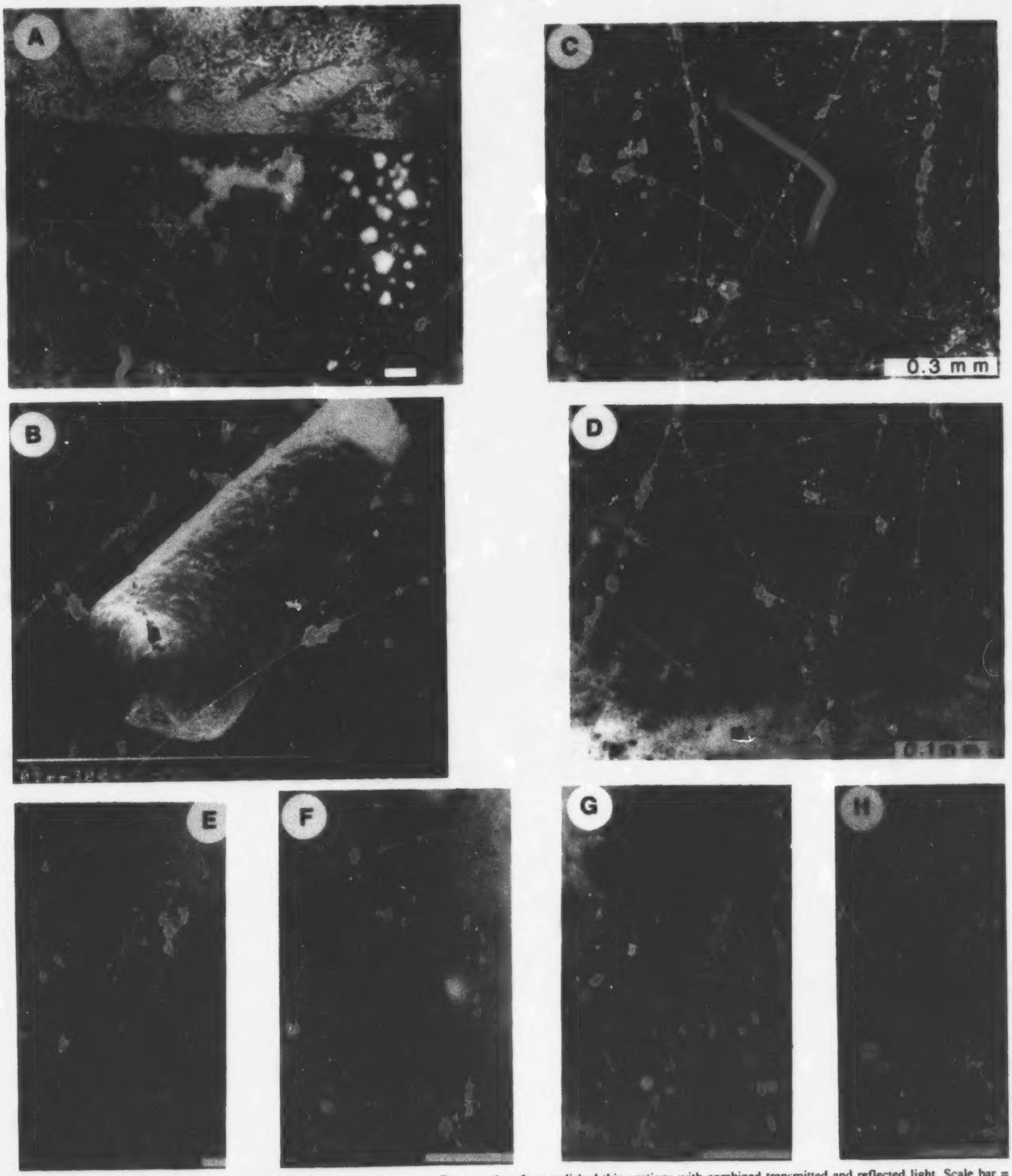

Figure 5: Pyrite textures of the Icebox Member. All photographs except B were taken from polished thin sections with combined transmitted and reflected light. Scale bar = 0.01 mm except where otherwise indicated.

- A: section of chitinozoan filled with pyrite.
- B: S.E.M. photograph of chitinozoan filled with pyrite. Note the "bumpy" outer surface produced by impression of pyrite crystals, and crystals visible through the fracture at the base of the test.
- C: "thread"-like pyrite and pyrite infills.
- D: pyrite filling the cavity of a scolecodont and replacing an organic filament.
- E: pyrite replacement of a nostocacean filament.
- F: framboidal pyrite replacing spherical algal or bacterial cells.
- G and H: cylindrical and prismatic pyrite crystals replacing nostocacean filaments.

(6) abraded pyrite grain of detrital origin;
(7) reworked fragment of pyrite oöid; and
(8) ill-defined nucleus.

Work in progress (Frizzo and Binda, 1989) has revealed an important mode of cortex accretion which was not detected by Binda and Simpson (1989). Some oöids (*e.g.* Fig. 4D) clearly show the incorporation of framboids in the formation of the cortex. Clusters of partially recrystallized framboids have also been recognized at the nucleus of some pyrite-coated grain. Given the ease with which framboids tend to recrystallize, and thus escape detection in un-etched polished sections of coated grains, it is suspected that this process may play a major role in cortex formation.

REPLACEMENT AND INFILL

Some samples of the Icebox Member contain up to 20 percent (estimated) pyrite, mostly as thread-like replaced organic fragments and as infill of microfossils (Fig. 5).

The acicular and filamental structures (Fig. 5C) up to a few hundred micrometres in length, are heterogeneous fragments, probably representing a variety of organic elements such as spines, spiculae, and algal or bacterial filaments. Some filaments can safely be attributed by cyanobacteria (Cyanophyta). The chains of pyrite "beads" of figure 5E, G, and H bear remarkable resemblance to Nostocaceae, particularly to modern genera *Anabaena*, *Anabaenopsis*, and *Nostoc*. The chain of framboids of Figure 5F is also probably the result of pyrite filling algal or bacterial cells. In all likelihood, framboids formed where spherical cells were available (Rickard, 1970), whereas single crystals formed in cylindrical cells.

Pyrite-replaced bacterial and algal filaments which are virtually identical to the ones of the Icebox Member have been described by Wicander and Schopf (1974) from the Early Devonian Kalkberg Limestone of New York State, and interpreted as a result of pyrite infilling while the cell wall was still intact.

Small shelly fossils, probably brachiopods, completely replaced by a mosaic of crystalline pyrite have been observed in the Icebox Member. In sandstone of the Black Island, a few burrows filled with pyrite pseudomorphs after detrital silicate grains have been noted. Also a large number of chitinozoans (Fig. 5A and B) and a few scolecodonts (Fig. 5D) of the Icebox Member are filled with pyrite.

Chitinozoans are organic-walled, bottle-shaped microfossils of uncertain biological affinity. Traditionally they have been placed within the Protozoa, but it has been argued that they may be metazoan eggs or capsule, or reproductive bodies of graptolites (Jenkins, 1970). Pyrite crystals a few micrometres in diameter occur inside the cavity producing distinct protuberances on the outer surface of the test (Fig. 5B). The impressions were probably produced as a result of differential yield of the rigid pyrite and of the more pliable test during compaction. This suggests a pre-compaction formation of the pyrite. Figure 5A and B clearly show that chitinozoan tests are not replaced by pyrite; possibly their composition made them resistant to the extensive pyritization that seems to have affected most organic material in the Icebox Member. Jenkins (1970) reported chemical analyses showing that chitinozoan tests are not true chitin, but rather a glycoprotein containing amino acids such as glutamic acid, aspartic acid, and others.

Scolecodonts are dispersed jaws of marine polychaete worms of chitinous or cuticulinous composition (Jansonius and Craig, 1971). They are fairly common microfossils of the Icebox Member and their pulpal cavity (myocoele) is commonly infilled with pyrite crystals (Fig. 5D) whereas their solid parts do not appear to be affected by pyritization.

SULPHIDE IN THE MATRIX OF FE-IRONSTONE

Thin layers of oölitic ironstone have been recorded from a few drillholes in the Icebox Member (Paterson, 1971). In the TW Pan-Am Whitebear Cr drillhole of Figure 2, geothite oöids up to 2 mm in diameter occur sparsely in the shale over about two metres of section, becoming the main constituent of the rock in a few bands up to 15 cm thick. Detrital quartz is scarce in these oölitic rocks.

Within the shale matrix, pyrite occurs as: 1) euhedral crystals 2 to 20 μm in diameter; 2) framboids, isolated or in clusters, the latter filling cavities of probable organic origin; 3) acicular or thread-like arrangements of small crystals; 4) small crystals infilling chitinozoan tests; and 5) larger masses, up to 0.5 mm in diameter, intergrown with chalcopyrite (Fig. 4E and F).

Examination of a few polished thin sections in ultraviolet light microscopy revealed the presence of abundant un-oxidized organic matter of algal or bacterial origin within the shale. Although a discussion of the genesis of oölitic ironstones (see Maynard, 1983) is beyond the scope of this paper, it is obvious that the formation of goethite must have taken place in the presence of a higher level of concentration of oxygen than precipitation of pyrite, and that preservation of organic matter requires similar low oxygen conditions required for sulphide precipitation. It seems logical, therefore, to attribute the formation of oxide and sulphide in the same rock to two distinct time phases: goethite during oxic sedimentation, sulphide during anoxic diagenesis.

AMINO ACIDS OF SULPHIDE-COATED GRAINS

Three hand-picked concentrates of sulphide-coated grains from the Black Island Member were analyzed for amino acids at the Department of Agrarian Biotechnologies of the University of Padova, Italy. Total amino acid content of the grains ranges between 9.5 and 26.1 μmal/g. Considering that modern carbonate oöids from the Bahamas average 4.5 μmal/g and Jurassic carbonate oöids from the Eastern Alps average 0.14 μmal/g, the amount of amino acids in the sulphide-coated grains is extremely high and has been interpreted by Binda *et al.* (1991) as evidence of the active role played by bacteria in the formation of the oöids and also of the excellent preservation of the organic content within the textures.

Glycine, leucine, glutamic acid, and aspartic acid are the most abundant amino acids present. The relative abundance of the various types of amino acids in the samples is probably a function of the types of bacteria that mediated the precipitation of sulphide, of the substrate on which the bacteria were feeding, of the products of complex biosynthetic reactions, as well as of preservation during post-depositional history of the sediment.

The presence of non-racemized amino acids in the sulphide-coated grains raises the problem of possible contamination by modern organic matter, since optically-active amino acids are not normally found in geological material older than the Neogene. The problem has been discussed by Kvenvolden *et al.* (1969) in the context of non-racemized amino acids occurring in Precambrian chert. Laboratory contamination can be excluded in the case of the sulphide-coated grains since only intact oöids were selected, great care was taken in all procedures, and blanks were employed. However, the possibility of *in situ* contamination cannot be totally discounted.

In favour of a syngenetic origin of the amino acids is the relative high content of methionine (4.2 percent in sulphide-coated grains *vs.* 1.4 percent in Bahaman oöids and 0.6 percent in Jurassic carbonate oöids). As stated by Krouse and McCready (1979, p. 316) "most micro-organisms assimilate sulfate sulfur which they reduce to the level of sulfide and incorporate into the sulfur-containing metabolites such as cysteine, methionine and gluthathione". Thus the abundance of methionine suggest preservation of organic matter that was present at the time of sulphide formation. Organic matter introduced at a later date would not necessarily include S-containing amino acids. Tests aimed at solving this problem are continuing at the University of Padova.

SULPHUR ISOTOPES

Three concentrates of sulphide-coated grains analyzed by I. Lambert at the Baas Becking Geobiological Laboratory (Canberra, Australia) gave the following $\delta^{34}S$ ‰ : +12.8, +3.3, +1.7 (Binda and Simpson, 1989). The values were interpreted to imply bacterial sulphate reduction a short distance below sea floor, in sediments in which influx of sea water was limited.

In a study of Jurassic shales Fisher and Hudson (1987) concluded that sulphate reduction can take place either in an open system in an anoxic water column in which supply of sulphate is unlimited, or in a partially closed system by diffusion of sulphate. In an open system the difference between the sulphur isotopic ratio of the pyrite and that of coeval sea water sulphate is very negative ($\delta^{34}S$ up to -50 ‰) whereas in a partially open system the difference ranges up to -25 ‰.

The $\delta^{34}S$ of Late Ordovician sulphates in equilibrium with sea water is given by Claypool *et al.* (1980) as approximately -28 ‰. Thus, the $\delta^{34}S$ of sulphide-coated grains of the Winnipeg Formation (-15.2, -24.7, -26.3 ‰), averaging -22.1 ‰, is compatible with the partially closed system of the restricted facies (*e.g.* Lower Oxford Clay) of Fisher and Hudson (1987).

DISCUSSION

Diagenetic precipitation of pyrite in sediments is generally interpreted as the result of reaction between H_2S produced through bacterial reduction of sulphate according to a reaction of the type: $2CH_2O + SO_4^{2-} \dashrightarrow H_2S + 2HCO_3^-$ (Westrich and Berner, 1984); and iron which may be present, in the interstitial waters as bacteriogenically-bound colloidal species such as $[Fe(H_2O)_6]^{2+}$ or $[Fe(H_2O)_4(OH)_2]^+$ (Ferris *et al.*, 1987), or in the sediments as Fe oxides and hydroxides.

Bacterial reduction of seawater sulphate can be performed by several types of S-reducing dissimilatory anaerobes, such as *Desulfovibrio*, *Desulfotomaculum*, and *Desulfomonas* whose activity leads to the oxidation of available organic matter according to complex biochemical reactions. However, reduced sulphur assimilated by bacteria, algae, and fungi may also be a source of sulphur for sulphide mineralization.

Binda and Simpson (1989) have presented petrographic evidence for early diagenetic origin of the pyrite-coated grains, these grains did not form by replacement of pre-existing oöids. A bacterially-mediated precipitation of sulphide is indicated by the sulphur isotopic ratios and is corroborated by preliminary amino acids results.

Berner (1981) showed that, in sediments, pyrite forms in anoxic sulphidic environments in which oxygen concentration is $<10^{-6}$ mole/litre and $H_2S \geq 10^{-6}$ mole/litre. Such environments exist (1) in the bottom sediments of euxinic basins characterized by undisturbed laminated muds, or (2) below the water/sediment interface in organic-rich bioturbated sediments. The sediments of the Winnipeg Formation show a considerable degree of bioturbation and therefore it is unlikely that anoxic conditions existed at the water/sediment interface. As discussed in the previous section of this paper, sulphur isotope ratios indicate precipitation of pyrite in a partially closed system.

Further evidence that the water/sediment interface was not anoxic during deposition of the Icebox Member is provided by the goethite oöids. The presence of hydroxide oöids and diagenetic pyrite in the same sedimentary rock has implications on the much debated question of ironstone genesis. Kimberley (1979) proposed that oölitic ironstones form by diagenetic replacement of carbonate oöids. This diagenetic model was questioned by a number of workers (*e.g.* Moltzer and Binda, 1988). Clearly, the Fe-hydroxide oöids of the Icebox Member could not have formed in a diagenetic environment in which not only was pyrite being precipitated, but also organic matter was not being oxidized.

The origin of framboids has been the subject of debate since Schneiderhohn's (1923) suggestion that they form in bacterial colonies, with each crystallite infilling a bacterial cell. The debate is summarized by Rickard (1970) who proposed that the spheroidicity of framboids requires "pseudomorphism of a pre-existing spherical body" (*ibid.*, p. 291). While accepting the possibility that framboids may form inside spherical unicellular micro-organisms, Rickard (1970) proposed that the majority of framboids in marine sediments form by replacement of organic globules such as oil droplets, or of gas vacuoles. Rickard envisaged precipitation of pyrite as a two-stage low-temperature process starting with sulphidization of goethite to form mackinawite:

$$2\,FeOOH + H_2S = 2\,FeS + S + OH^- + 3\,H_2O$$

followed by a reaction between mackinawite and polysulphide ions:

$$FeS + S_2^{2-} = FeS_2 + S^{2-}.$$

Rickard (1970) further stated that this polysulphide reaction differs from the solid state sulphur oxidation of mackinawite to give marcasite. Obviously both processes were operative during early diagenesis of the Winnipeg Formation since marcasite rosettes and pyrite framboids must have formed penecontemporaneously, as both may be surrounded by a pyrite cortex (Fig. 4B and D).

In the Winnipeg Formation, framboids occur either isolated or in clusters and chains. Possibly the two latter types represent replacement of spherical coccoidal micro-organisms. Framboids have not been observed in the thread-like chains formed by infill of cylindrical cells of nostocacean cyanobacteria (Fig. 5 E,G,H). A mode of framboid formation involving early diagenetic replacement of either oil droplets or gas vacuoles is entirely possible. Agglomerations of cup-like structures of the cyanobacterium *Gloeocapsomorpha prisca* forming an "alginite" are abundant in the Icebox Member (Osadetz *et al.*, 1989). Aizenshtat *et al.* (1984) have shown that up to 7.65 mg/g of free hydrocarbon are present in the top 30 cm of sediment in Solar Lake, Sinai, where microbial mats are presently forming. Gas vacuoles can be expected to have formed during early diagenesis of the Winnipeg Formation as a result of fermentation.

However, an entirely inorganic origin of framboids cannot be excluded: Berner (1969) synthesized pyrite framboids in sterile aqueous medium at T=65°C and pH=7.9 using H_2S, $FeSO_4$, and orthohombic sulphur. Black FeS precipitated immediately and pyrite framboids formed within two weeks.

MODERN ANALOGUES

Modern environments in which framboids are forming have been reported by a number of authors (*e.g.* Love, 1967). Well-developed sulphide-coated grains have not been observed in modern sediments. However, textures which may be considered precursors of the oöids of the Black Island Member have been reported from at least two localities.

In the top 10 cm of sandy muds of the Venice Lagoon, in addition to framboid clusters filling tests of micro-organisms, incipient pyrite cortices can be observed forming around partially replaced detrital grains of Fe-oxide and Fe-Ti-oxide (Frizzo and Binda, 1989).

Canfield and Berner (1987) have described magnetite grains partially replaced by pyrite and completely coated by a cortex of pyrite spherules from the sediments of Long Island Sound. Below the zone of bioturbation, at a depth of 20-30 cm, many magnetite grains have pyrite overgrowths. At a depth of 60 cm, nearly all magnetite grains are enclosed in a pyrite shell.

Recently, Lin and Morse (1991) have shown a progressive increase in the degree of pyritization down to a depth of 4.5 m in sediments of the Mississippi Delta and of the Gulf of Mexico continental shelf and slope. Most of the pyrite is formed on the shelf. Concentration of organic matter is the limiting factor for sulphate reduction and pyrite precipitation.

CONCLUSIONS

Petrographic, isotopic and amino acid analyses indicate that the pyrite textures of the Oruovician Winnipeg Formation of Saskatchewan formed during early diagenesis in a partially closed system, probably a few to several tens of centimetres below the water/sediment interface. Sulphide precipitation was mediated by bacterial reduction of sea-water sulphate. Similar pyrite grains are forming in modern sediments.

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LARGE SCALE FLUID MOVEMENT IN THE WESTERN CANADIAN SEDIMENTARY BASIN AS RECORDED BY FLUID INCLUSIONS IN EVAPORITES

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ABSTRACT

After primary migration out of source beds, further migration of petroleum is not achieved by its buoyancy alone but also by movement of aqueous fluids in the porous rock through which the petroleum migrates. Temperatures required for the generation of oil and the hydrodynamic conditions during secondary and tertiary migration, therefore, influence the ultimate trapping site for petroleum.

Times of fluid movement have been documented in the Devonian Prairie Evaporite by vertical extent of collapse features, characterization of the properties of fluids in inclusions in halite, and radiometric dating of evaporite minerals. Halites from the potash ore zones of the Prairie Evaporite in the Saskatoon area were recrystallized by migrating fluids and contain in their fluid inclusions a record of the temperature and stable isotopic compositions of the fluids. Three events can be recognized. The first, which occurred shortly after deposition of the Prairie Evaporite, was at a temperature of about 35°C (95°F), and involved water with δD near -92 and $\delta^{18}O$ near -8 which are typical of groundwaters from mid-latitudes. Two later events, at about 50°C (120°F) and 80°C (180°F), were important with respect to petroleum migration because of the substantial amount of fluid involved, the maturity of the basin during these events, and the effect of these fluids on the permeability of the reservoir rocks. The lower temperature event involved water with δD near -70 and $\delta^{18}O$ near -6, similar in isotopic composition to evolved basin fluids. The higher temperature event involved water with δD near -141 and $\delta^{18}O$ near -15 and occurred during maximum burial at the end of the Cretaceous. The stable isotopic compositions of these fluids are similar to brines presently residing within formations overlying the Prairie Evaporite but the temperatures were higher because the basin was deeper and the geothermal gradient was greater.

INTRODUCTION

Aqueous fluids migrating in sedimentary basins are important carriers of dissolved metals and petroleum and often control the formation of exploitable mineral and oil deposits (Garven, 1985; Chapman, 1987; Ge and Garven, 1989; Bethke and Marshak, 1990; Cumming *et al.*, 1990). Reactions between the migrating fluid and the various mineral phases encountered in the basin affect the porosity and permeability of potential reservoirs through dissolution, recrystallization, and replacement by new phases. The extent of all these reactions are determined by the pressure, temperature, and chemical characteristics of the sediment and fluid (Kharaka and Carothers, 1986). For example, migration of warm brines from deep basin levels results in anomalously high temperatures recorded by organic matter in sediments (Bethke and Marshak, 1990). Fluid migration also may effect secondary and tertiary petroleum migration (Bethke and Marshak, 1990).

Migration of fluids in sedimentary basins can be induced by compaction of water-saturated sediments shortly after deposition, or by thrusting, compression, and tectonic uplift in adjacent orogens (Garven, 1985; Garven, 1989; Ge and Garven, 1989; Bethke and Marshak, 1990). Bethke and Marshak (1990) concluded through modelling that tectonic uplift resulted in the greatest fluid flow events.

The Western Canada Sedimentary Basin has been subjected to episodes of tectonic activity producing topographic uplift during Lower Devonian, Upper Mississippian, Lower Pennsylvanian to Jurassic time, and also during formation of the Cordillera from Middle Jurassic through Miocene time (Porter *et al.*, 1982). Expulsion of water by sediment compaction in the Late Pennsylvanian (295 Ma) has been credited with formation of the Pine Point Pb-Zn deposit (Cumming *et al.*, 1990). Later fluid migration due to Cretaceous uplift has been postulated as the mechanism for the formation of massive oil sands deposits in Alberta (Garven, 1989). Fluid migration events in the Western Canada Sedimentary Basin also have resulted in dissolution and recrystallization of the evaporite minerals of the Prairie Evaporite of southern Saskatchewan (McTavish and Vigrass, 1987; Chipley and Kyser, 1990).

The Prairie Evaporite is a Middle Devonian sequence about 200 m (660 ft) thick in the area near Saskatoon, Saskatchewan, (Fig. 1) and is presently buried to about 700 m (2,300 ft) along its northern edge and to 3,650 m (11,980 ft) in North Dakota (Fig. 2; Fuzesy, 1982). The sequence is composed of a lower portion of alternating halite and anhydrite, overlain by about 70 m (230 ft) of halite containing four potash ore horizons named, in ascending stratigraphic order, the Esterhazy, White Bear, Belle Plaine, and Patience Lake Members (Fig. 2). The ore consists of varying proportions of sylvite (KCl) and carnallite ($KMgCl_3 \cdot 6H_2O$) intermixed with halite ($NaCl$), interrupted locally by leach and collapse deposits. Leach anomalies are recognized by the partial or total replacement of the ore minerals by halite whereas, in collapse anomalies, the evaporite has been partially or completely removed by dissolution, thereby allowing collapse of overlying rocks into the voids to form a breccia with a halite matrix (DeMille *et al.*, 1964; McIntosh and Wardlaw, 1968). The dating of collapse and sagging of overlying beds into solution cavities allows determination of the ages of dissolution and, therefore, timing of fluid migrations (McTavish and Vigrass, 1987). The last correlate broadly with tectonic events defined by Porter *et al.* (1982).

Recrystallization of the evaporite minerals resulted in entrapment of post-Middle Devonian migrating fluids as inclusions in halite. Halites are, therefore, not only a record of fluid migration events, but also include a sample of the migrating

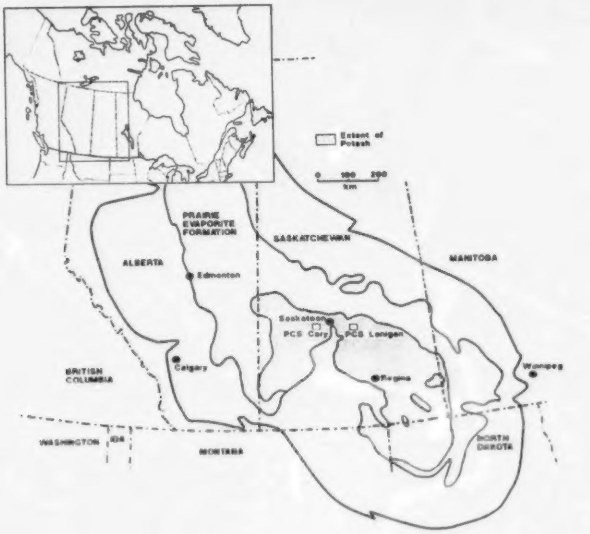

Figure 1: Map showing the location of the PCS Cory and PCS Lanigan potash mines in Saskatchewan, and the areal extent of potash (ruled area) in the Prairie Evaporite of the Western Canada Sedimentary Basin.

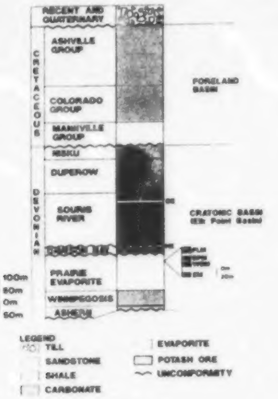

Figure 2: Stratigraphic section through the Western Canada Sedimentary Basin in the Saskatoon area. EM - Esterhazy Member, WBM - White Bear Member, BPM - Belle Plaine Member, PLM - Patience Lake Member, HE - Hubbard Evaporite, DE - Davidson Evaporite.

fluid itself. The temperatures at which the fluids were trapped and the chemical and stable isotopic compositions of the fluids are valuable in determining both fluid source and type of reactions that occurred.

Ore samples of the Patience Lake Member were obtained from the Potash Corporation of Saskatchewan (PCS) Cory and Lanigan mines in order to determine the number of generations of fluids and the temperatures, chemical and isotopic compositions of the fluids. The purpose of this study is to acquire a better understanding of the fluid history of the Western Canada Sedimentary Basin and possible relationships to oil and mineral deposits therein.

ANALYTICAL METHODS

Halite crystals were removed from ore samples and cleaved to give thin (<1 mm) slabs for microscopic examination. Generations of fluid inclusions were distinguished through textural differences, chemical compositions, and daughter mineral dissolution temperatures which were observed using a USGS-type fluid inclusion stage. The chemical composition of the fluids was determined by identification of the daughter minerals and melting characteristics of the frozen fluids. Daughter mineral dissolution temperatures (Tm-dm) were obtained by heating the fluid inclusions until sylvite and carnallite daughter minerals dissolved. Extraction and analysis of the hydrogen and oxygen isotopic compositions of the fluid inclusion fluids were carried out using the method of Koehler *et al.* (1991).

Stable isotopic compositions of hydrogen and oxygen in water and brine are measured relative to an international standard, designated Standard Mean Ocean Water (SMOW) and are reported in units of per mil using delta notation, and defined for oxygen as:

$$\delta^{18}O = [(^{18}O/^{16}O)_{sample} - (^{18}O/^{16}O)_{standard}/(^{18}O/^{16}O)_{standard}] * 1000$$

A similar equation can be written for the two stable isotopes of hydrogen, deuterium (D) and hydrogen (H). By these definitions, SMOW has δD and $\delta^{18}O$ values of zero per mil.

When seawater evaporates to form atmospheric vapour, H and ^{16}O escape preferentially. This enrichment of water vapour in H and ^{16}O is enhanced by distance from the ocean, higher altitude, and lower mean annual temperature, *i.e.*, higher latitude (Kharaka and Carothers, 1986). Precipitation, resulting from condensation of atmospheric water vapour, is enriched in D and ^{18}O relative to

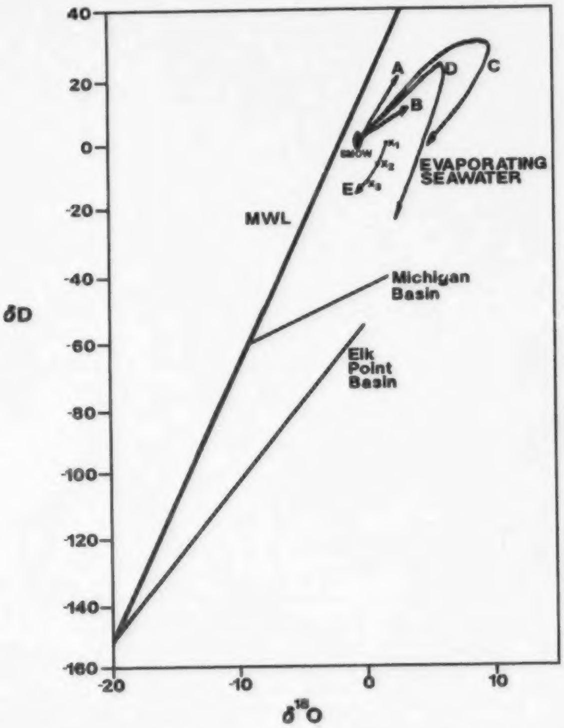

Figure 3: δD and $\delta^{18}O$ values for meteoric water (MWL), seawater (SMOW), basin brines, and changes in the isotopic compositions of seawater during evaporation. A - humid evaporation conditions, B - arid evaporation conditions (Holser, 1979), C - evaporation studies in experimental and natural systems (Holser, 1979), D - results from Ojo De Libre lagoon, Mexico, evaporated to halite precipitation (Pierre *et al.*, 1984), E - and evaporated seawater from salt crystallizing ponds, Bonaire Island, X_1 = [illegible]-gypsum, X_2 = pre-halite, and X_3 = post-halite, pre-sylvite.

the water vapour, but remains depleted relative to seawater (Welhan, 1987). A plot of hydrogen and oxygen isotopic compositions of precipitation define the meteoric water line (MWL) (Fig. 3). The extreme evaporation of seawater required to precipitate evaporites first results in enrichment of the remaining brine in D and ^{18}O due to loss of H and ^{16}O to the vapour phase, but with continued evaporation the trend reverses. The remaining brine becomes progressively depleted in D and ^{18}O as shown in the hooked trajectory of δD vs $\delta^{18}O$ in the graph of Figure 3.

Fluids in sedimentary basins have isotopic compositions that plot on lines with slopes lower than the meteoric water line (Fig. 3). Basin lines intersect the meteoric water line at approximately the isotopic composition of local modern meteoric water and the fluids usually become progressively enriched in D and ^{18}O with depth (*e.g.* Kharaka and Carothers, 1986).

FLUIDS IN THE WESTERN CANADA SEDIMENTARY BASIN

By microscopic examination of cleavage fragments of halite two distinct textures formed by fluid inclusions are revealed. Some occur in concentrated bands along halite growth zones, separated by bands with few inclusions. This is called chevron halite. Other fluid inclusions are distributed randomly. Where both types of inclusions are present in the same crystal, the chevron texture always occurs at the center of the crystal and is surrounded by the randomly distributed inclusions. Therefore, the chevron fluid inclusions contain fluids older than those randomly distributed.

Fluids that leak into potash mines and recrystallize halite quickly reach chemical equilibrium by dissolution of evaporite minerals (Zuber, 1978; Wittrup and Kyser, 1990; Chipley and Kyser, 1990) and thus are saturated with the evaporite phases in the environs of the halite at the time of trapping. Chemical compositions of the fluids in the inclusions, therefore, do not represent the composition of the source of the fluids but rather those of the dissolved and recrystallized evaporites along their migration path. Halite fluid inclusions from PCS Cory contain sylvite (KCl) daughter minerals and those from PCS Lanigan dominantly include carnallite ($KMgCl_3 \cdot 6H_2O$) daughter minerals (Fig. 4.), similar to the mineralogy of the respective ore horizons. The melting characteristics of the fluids in halite fluid inclusions from both mines in the Patience Lake Member indicate that the fluids are a Na-K-Ca-Mg-Cl brine, although fluid inclusions in the PCS Lanigan mine are richer in magnesium as indicated by saturation with carnallite. Magnesium-rich fluids have been proposed as agents in the dolomitization of limestones (Gregg and Shelton, 1989), and some dolomitizing fluids may have had their origins in the dissolution and recrystallization of evaporite minerals from the Prairie Evaporite.

Minimum trapping temperatures of the fluids can be determined by redissolving the sylvite and carnallite daughter minerals. Daughter mineral dissolution temperatures should be equivalent to true trapping temperatures, because dissolution of solid phases involves volume changes that are small relative to homogenization of liquid and vapour phases, so that pressure corrections should be small. Daughter mineral dissolution temperatures from fluid inclusions in halite range from 35°C (95°F) to 110°C (230°F) and are distributed bimodally, indicating that at least two different fluids were trapped as a result of dissolution and recrystallization (Fig. 4). Daughter mineral dissolution temperatures in the chevron fluid inclusions range from 35°C (95°F) to 50°C (120°F) and in the randomly distributed fluid inclusions from 50°C (120°F) to 110°C (230°F). These are greater than the present temperature of about 27°C (80°F) of the Patience Lake Member in the Saskatoon area. To attain the temperatures indicated by the fluid inclusions, the Prairie Evaporite must have been buried deeper, or else the geothermal gradient was greater.

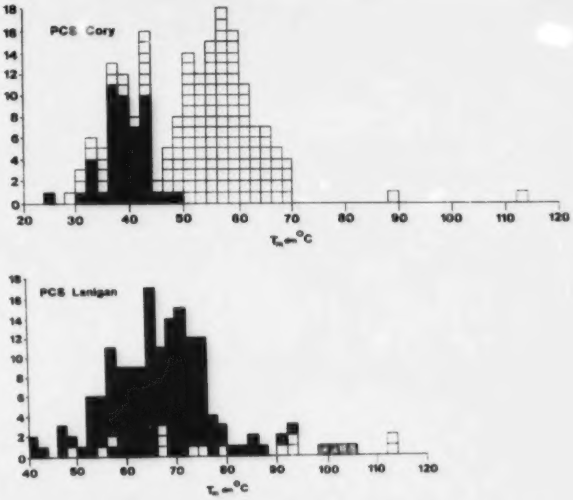

Figure 4: Histogram of fluid inclusion daughter mineral dissolution temperatures (Tm-dm) for halites from a) PCS Cory and b) PCS Lanigan. Closed squares are temperatures from chevron fluid inclusions and cross-slashed and open squares for carnallite and sylvite daughter minerals, respectively, in randomly distributed fluid inclusions.

The chevron fluid inclusions have δD values near -92 and $\delta^{18}O$ values near -8 (Fig. 5), which do not approximate the stable isotopic compositions of evaporated seawater, residual brine from seawater evaporation, or water expelled by compaction of marine sediments. An isotopically depleted water component similar to meteoric water, as well as reaction with sediments may have been involved to produce the values indicated. A single fluid event produced the chevron halites and this event occurred shortly after deposition of the Prairie Evaporite, most likely during Late Devonian or Early Mississippian time when the area was at low latitudes and being eroded.

The values determined from the randomly distributed fluid inclusions indicate a low temperature event near 50°C (120°F) and a higher temperature event near 80°C (180°F) (Chipley and Kyser, 1990). The lower temperature event fluid has δD near -70 and $\delta^{18}O$ near -6, similar in isotopic composition to the most evolved basin fluids in the Williston Basin and to fluids presently in the Dawson Bay Formation. The high temperature fluid has δD near -141 and $\delta^{18}O$ near -15. These are values similar to those of modern meteoric water and of fluids within the overlying Cretaceous aquifers. The similarity of the isotopic composition to the fluids in the Cretaceous aquifers and the high dissolution temperature probably indicate that this recrystallization event occurred before the topographic erosion near the end of the Cretaceous. Access of the fluids to the Prairie Evaporite from overlying aquifers may have been through fractures caused by reactivation of deep structures during tectonic events (Dunn, 1982).

SYNOPSIS

Fluid inclusions in halite from the Patience Lake Member of the Prairie Evaporite contain a valuable record of post-Middle Devonian fluid migrations in the Western Canada Sedimentary Basin. The temperatures of the fluids, as indicated by daughter

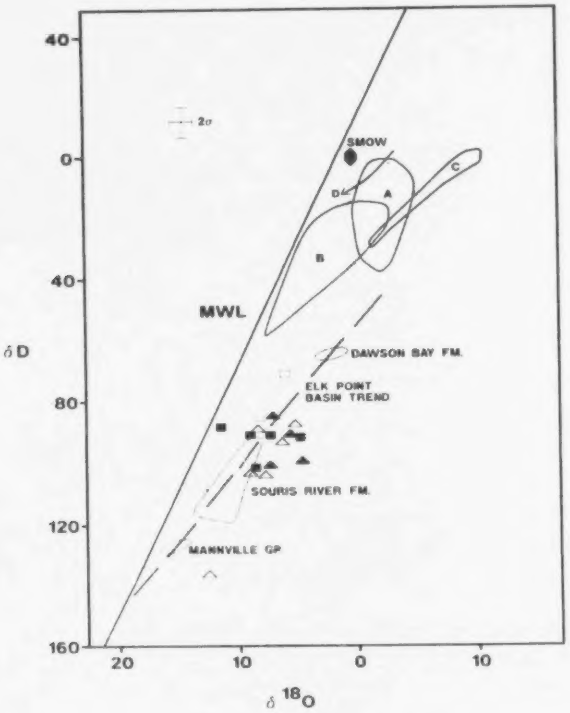

Figure 5: Stable isotopic compositions of water from formations overlying the Prairie Evaporite (Elk Point Basin Trend) (Wittrup and Kyser, 1990) and fluids extracted from fluid inclusions. Fields A and B represent halite fluid inclusion waters from chevron and non-chevron fluid inclusions, respectively, in the Palo Duro Basin (Knauth and Beeunas, 1986), field C represents water in halite fluid inclusions from the Delaware Basin thought to be derived from dehydration of gypsum which recrystallized the halite (O'Neil *et al.*, 1986) and curve D represents evaporation of seawater from Bonaire Island salt crystallization ponds. Squares (PCS Cory) and triangles (PCS Lanigan) represent fluids extracted from halite fluid inclusions from the Patience Lake Member of the Prairie Evaporite.

mineral dissolution temperatures, are higher than the present temperature of the Prairie Evaporite and may be due to deeper burial of the evaporite or to greater geothermal gradient in the past.

The stable isotopic compositions of the fluids reveal that the fluids that most affected the Western Canada Sedimentary Basin are complex mixtures of meteoric waters, evolved basin brines, and seawater. The presence of pure seawater or evaporated seawater was not indicated in any halite fluid inclusions. Chemical compositions of fluid inclusions represent fluids that dissolved and recrystallized evaporites along their migration paths. Some of these fluids are magnesium-rich and may have been responsible for dolomitization of limestones elsewhere in the basin resulting in increased porosity and permeability.

Fluid migration events in the Western Canada Sedimentary Basin were probably induced by tectonic activity. Reactivated deep structures probably provided access for the fluids to the Prairie Evaporite.

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FLUID EVENTS AS RECORDED IN THE HYDRATION WATER OF CARNALLITE FROM THE ELK POINT BASIN OF WESTERN CANADA

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ABSTRACT

The stable isotopic compositions of hydration water in carnallites ($KMgCl_3 \cdot 6H_2O$) from the Esterhazy Member of the Prairie Evaporite indicate that this mineral in its present form was not deposited during the evaporation of ancient seawater, but has been recrystallized and subjected to diagenetic processes involving fluids that have their origins in formations above the evaporite. Most of the carnallite now present as intercalations in the ore and as carnallite pods is the result of relatively recent fluid events at low temperature (25-40°C). The stable isotopic compositions of hydration waters in carnallites collected from potash mines at Rocanville and Esterhazy, Saskatchewan, indicate that there have been at least three different fluid events responsible for the formation of carnallite, whereas samples from a drill-hole core near Lanigan indicate carnallite formed from a number of distinct fluid events. Much of the carnallite in this evaporite deposit may be the result of subsurface, post-depositional processes involving the reaction of sylvite (KCl) with Mg-rich basinal brines. These Mg-rich brines may be related to de-watering of primary carnallite during early burial of the potash deposit and may have reacted with limestones in this basin to produce dolomites, therefore affecting the porosity of these potential reservoir rocks.

INTRODUCTION

The origin of sedimentary brines and the timing of brine movement through sedimentary basins have remained problems of modern sedimentary and petroleum geology. It is evident that pervasive migration of brines occurs within sedimentary sequences, as revealed by the presence of ore deposits and the distribution of petroleum in sedimentary basins (see Bethke and Marshek, 1990, for review). However, despite decades of research, the origin and timing of fluid flow through basin sediments have remained poorly understood.

Evaporite sequences within sedimentary basins provide an opportunity to examine diagenetic fluid events because evaporites, especially those which contain chloride minerals, are easily recrystallized by later fluids and therefore may contain a geochemical record of post-depositional alteration of the sedimentary package. In addition, brines that may be the result of interaction with evaporites have been proposed as the fluids responsible for dolomitization of carbonate units in the subsurface (Aulstead *et al.*, 1988; Machel and Anderson, 1989). Many evaporite sequences have been the focus of research on the origin of basin brines, including the Salado Formation of the Delaware Basin (O'Neil *et al.*, 1986; Horita *et al.*, 1991), the Palo Duro Basin, Texas (Knauth and Beeunas, 1986), and the Paradox Basin, Utah (Huff and Wampler, 1990).

Our research has been concerned with one of these ancient evaporite sequences, the Middle Devonian Prairie Evaporite of western Canada. There is abundant evidence that this evaporite has been recrystallized by later fluids (see for example Wardlaw, 1968, and Wardlaw and Watson, 1966). Therefore, examination of evaporite minerals in this deposit may provide direct evidence of the origin and timing of fluid events in the Elk Point Basin.

The stable isotopic compositions of minerals and fluid inclusions in minerals have been used to determine the origins of fluids in many basins (Knauth and Beeunas, 1986; O'Neil *et al.*, 1986). Of particular interest in the Prairie Evaporite is the hydrated saline mineral, carnallite ($KMgCl_3 \cdot 6H_2O$). Because the fractionation of the isotopes of hydrogen and oxygen between the hydration water of carnallite and the brine from which this mineral precipitates is known and because this mineral is easily recrystallized by fluids, the origin of any fluid responsible for carnallite formation and therefore fluid movement in the Elk Point Basin may be constrained.

REGIONAL FRAMEWORK AND GENERAL GEOLOGY

Middle Devonian sedimentation in western Canada occurred within an elongated trough that extended from northern Alberta, through central and southern Saskatchewan, and into the northern United States (Fig. 1). This basin, of which the Williston Basin was a part, was bounded by the Western Alberta Arch to the west, the Central Montana Uplift to the south, and Precambrian shield areas to the north and northeast (Fig. 1). Two other positive elements, the Peace River Arch and the Meadow Lake Escarpment divided the Elk Point Basin into three sub-basins: the Northern Alberta, Central Alberta, and Saskatchewan sub-basins. The Middle Devonian sea transgressed in stages over these elements from northwest to southeast, sequentially filling each sub-basin (Gorrell and Alderman, 1968). The Prairie Evaporite represents one of a number of evaporitic units in the Central Alberta sub-basin, and is a facies of the Muskeg Formation in the Northern Alberta sub-basin (Fig. 2). Potash salts occur only in the Prairie Evaporite of the Saskatchewan sub-basin.

The Prairie Evaporite lies at a depth of approximately 1000 m (3280 ft) in the Saskatoon area. It dips gently to the southwest, at approximately 2 to 8 m/km (10 to 40 ft/mi). The evaporite can be divided into three main subdivisions (O-Y): 1) the Whitkow Salt, which consists largely of halite with anhydrite layers; 2) the Shell Lake Gypsum; and 3) the Leofnard Salt, which contains the economically important potash members. The potash-bearing members are, in ascending order, the Esterhazy, White Bear, Belle Plaine, and Patience Lake (Fig. 2). A detailed description of the Prairie Evaporite can be found in Holter (1969).

The potash members are essentially composed of three chloride salts, halite (NaCl), sylvite (KCl), and carnallite ($KMgCl_3 \cdot 6H_2O$). Other constituents, usually present in small amounts, consist of clays, quartz, anhydrite, and dolomite (Mossman *et al.*, 1982). Small amounts of finely disseminated iron oxides (haematite and goethite) impart the characteristic pinky-red colour to the potash salts.

There are numerous salt dissolution features in the Prairie Evaporite, the most extensive of which is solution that occurred along the Swift Current Platform (Fig. 1). In addition, there are other, smaller scale, dissolution features that locally interrupt the evaporite and can result in collapse of overlying formations into the Prairie Evaporite. These features are convincing evidence that the salts have interacted with later fluids.

The Prairie Evaporite is conformably underlain by the Winnipegosis Formation, which consists of reefal and inter-reefal dolomites. The Dawson Bay Formation of the Manitoba Group unconformably overlies it. The lower portion of the Dawson Bay Formation is composed of a thin series of dolomitic shales, called the Second Red Beds. Above these shales, the formation consists of fossiliferous dolomites and limestones. A small halite unit, the Hubbard Evaporite, occurs locally in the uppermost part of the Dawson Bay (Dunn, 1983). Where the Prairie Evaporite is absent, the Second Red Beds rest directly on the Winnipegosis Formation (Gorrell and Alderman, 1968).

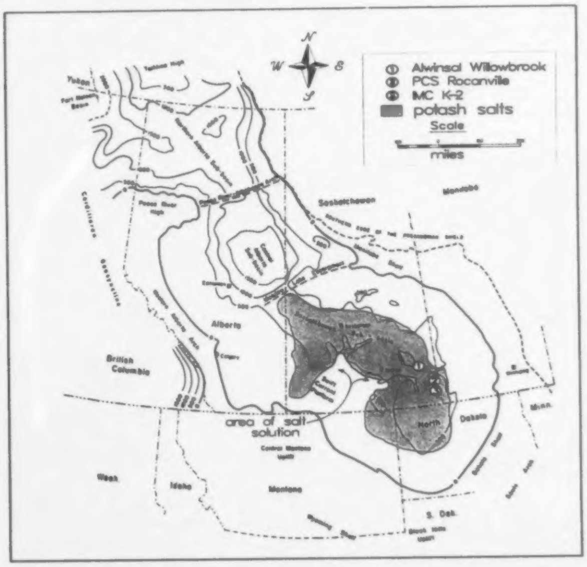

Figure 1: The Middle Devonian Elk Point Basin of western Canada. Isopach contours represent thicknesses of Elk Point sediments in feet. Shading represents the distribution of potassium salts in the Saskatchewan sub-basin (Modified after Holter, 1969). Also shown are the locations of PCS Rocanville and IMC K-2 potash mines and the Alwinsal Willowbrook drill hole.

ISOTOPIC PROPERTIES OF CARNALLITE

As early as 1947, it was recognized that there are differences in the stable isotopic compositions of coexisting phases as a result of small mass differences between the isotopes of elements incorporated into these phases (Urey, 1947). Isotopic fractionations of elements, such as hydrogen or oxygen will accompany chemical reactions, such as the precipitation of minerals from solutions or reactions of minerals with water, because of the different thermodynamic properties of isotopically different substances. Therefore, the stable isotopic compositions of fluids and minerals record their chemical history and temperatures, provided that the isotopic fractionations between minerals and fluids are known or can be estimated.

The fractionation of hydrogen and oxygen isotopes between the hydration water of carnallite and brine has been measured at different temperatures (Koehler *et al.*, 1990). At 25°C, laboratory experiments indicate that the hydration water of carnallite is depleted in D by about 40 per mil[1] relative to the brine, whereas at 41°C the hydration water of carnallite is only depleted by about 22 per mil relative to the brine. Oxygen isotopes show no fractionation at both 25 and 41°C. These isotope fractionations determined in the laboratory are consistent with fractionations measured between hydration water of carnallite and brine from carnallites precipitated in natural brine pools in potash mines (Koehler *et al.*, 1990).

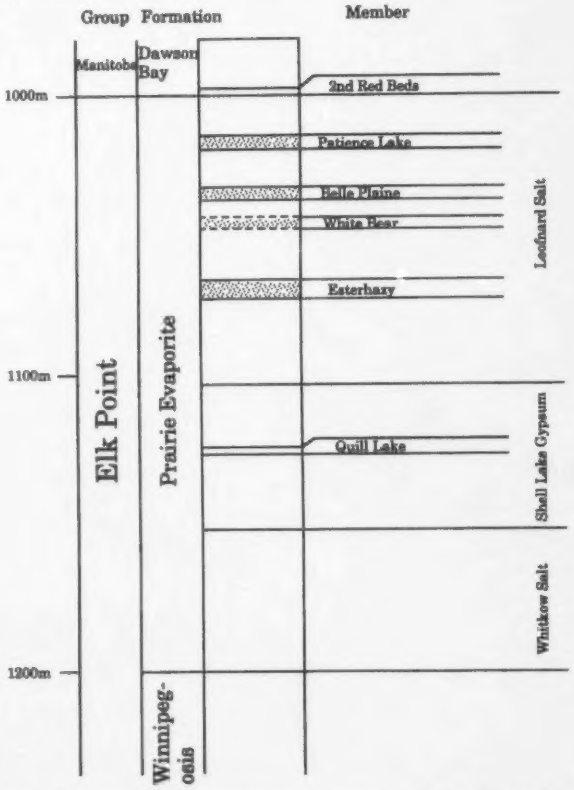

Figure 2: Simplified stratigraphy of the Prairie Evaporite. Stippled regions indicate the occurrence of potash salts.

STABLE ISOTOPIC COMPOSITION OF NATURAL CARNALLITES

Samples of carnallite were collected from the Esterhazy member of the Prairie Evaporite at two potash mines in Saskatchewan, IMC K-2 and PCS Rocanville (Fig. 1). Samples were taken from normal areas in the potash ore zone, from carnallite pods which occur sporadically throughout the potash members, and from an anomalous zone at PCS Rocanville.

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[1] stable isotopic compositions are reported in the W notation which is defined as $\delta(^{o}/oo) = [(R_{sample}/R_{standard})^{-1}] \times 1000$, where R is the D/H or $^{18}O/^{16}O$ ratio. The units of difference are in units of per mil ($^{o}/oo$), equivalent to differences in the ratios of parts per thousand. The standard is Standard Mean Ocean Water (SMOW).

Carnallite was also collected from the Esterhazy Member in the Alwinsal Willowbrook drill-hole core which was first sampled by Baadsgaard (1987) for Rb-Sr and K-Ca radiometric analysis.

The Alwinsal Willowbrook samples and the ore samples from PCS Rocanville and IMC K-2 have stable isotopic compositions that, with a few exceptions, lie on plots that parallel the Elk Point Trend (Fig. 3). The latter trend is plotted from the stable isotopic compositions of waters presently in formations above the Prairie Evaporite (Wittrup and Kyser, 1990). The stable isotopic composition of brines from which these carnallites formed can be estimated using the isotope fractionations measured in laboratory experiments discussed above. Because the estimated stable isotopic compositions of brines responsible for the precipitation of these carnallites follow the Elk Point Basin Trend, it appears probable that these carnallites were also formed from, or equilibrated with, waters which have their origin in formations above the potash deposits. None of these carnallites has a stable isotopic composition that suggests it was precipitated from evaporating seawater during initial formation of the evaporite.

The stable isotopic compositions of carnallites from the ore zone of PCS Rocanville and IMC K-2 indicate that carnallite formed at low temperatures from at least two fluids, one with δD values of -60 per mil and δ^{18}O values of about 0 per mil (A on Fig. 3), and the other with WD values of about -100 per mil and W^{18}O values of about -7 per mil (B on Fig. 3). B has an isotopic composition similar to some of the fluids presently leaking into mining levels at PCS Rocanville (δD=-100, δ^{18}O=-8), and A to fluids residing in the Dawson Bay Formation (δD = -60, δ^{18}O = -2) (Wittrup and Kyser, 1990). In contrast, the restricted range in the isotopic composition of the pod-type carnallites from PCS Rocanville and IMC K-2 indicates that these carnallites formed from a fluid similar to those presently within the Souris River or Duperow Formations (δD=-80, δ^{18}O=-6).

The isotopic compositions of fluids responsible for formation of carnallite from the anomalous region at PCS Rocanville indicate that these carnallites formed at low temperatures from a fluid having δD and δ^{18}O values of -40 and +4 per mil respectively (Fig. 3). Fluids of this stable isotopic composition do not exist in the basin today.

The plot of estimated stable isotopic compositions of fluids responsible for the formation of carnallites from the Alwinsal Willowbrook well also parallels the Elk Point Basin Trend. However, unlike the ore and pod samples from the potash mines, no distinct fluid events are recognizable (Fig. 3). These carnallites could have been precipitated, or equilibrated with, any of the fluids presently within the basin and probably represent repeated reworking of the evaporite deposit in this area by basin fluids. Radiometric dating of these carnallites and of carnallites from the Lower Zone of the Patience Lake Member at PCS Lanigan, near Lanigan, Saskatchewan, indicates 30 Ma (Baadsgaard, 1987). This suggests that at least some, and probably most, of the carnallites in the Alwinsal Willowbrook area are a result of relatively recent processes.

Carnallites having isotopic compositions which lie above those that formed at 25°C (Fig. 3), such as carnallite pods at PCS Rocanville and some of the ore and Alwinsal Willowbrook samples, most likely formed at either higher temperatures (*e.g.* >25°C) or from very small amounts of fluids such that all of the fluid was incorporated into the carnallite. In keeping with the isotopic properties of most substances, isotope fractionations between carnallite hydration water and brine are less at higher temperatures than at lower temperatures. Laboratory experiments indicate that hydrogen isotope fractionations between carnallite and brine at 41°C are only about half those at 25°C. Alternatively, if there is only enough fluid available to form carnallite, so that most of the water responsible for carnallite formation is incorporated into the hydration water of the mineral, the hydrogen isotopic composition of the carnallite will be identical to the brine. This latter method of carnallite formation could result by reaction of Mg-rich brines with pre-existing sylvite, rather than by direct precipitation from a brine that would occur if the fluid/rock ratios were high. The fractionation of stable isotopes of hydrogen is dependant on the amount of water incorporated into the hydration water of the mineral and, ultimately, the δD values of the carnallite will more closely resemble those of the parent brine. Laboratory experiments indicate that this process of carnallite formation at the expense of sylvite at low fluid/rock ratios can be duplicated. Because oxygen isotopes are not fractionated between carnallite hydration water and brine, this process does not affect the oxygen isotopic composition of the carnallites, *i.e.* the δ^{18}O values of carnallite always represent those of a precipitating brine, regardless of the temperature or fluid/rock ratios.

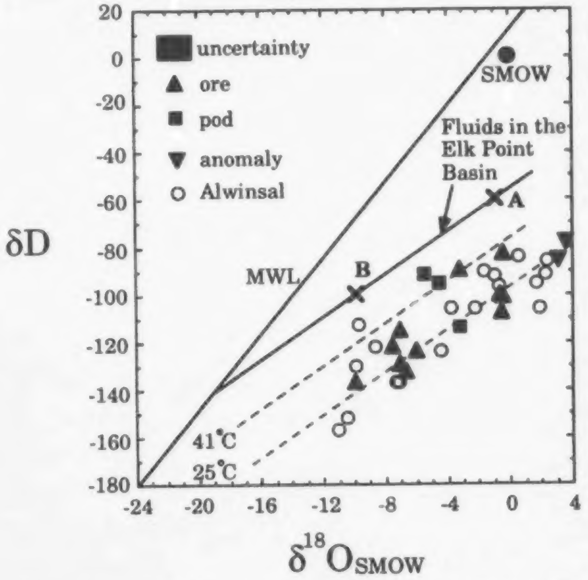

Figure 3: Stable isotopic compositions of hydration water in carnallite from PCS Rocanville and IMC K-2 potash mines (solid symbols), and the Alwinsal Willowbrook core (open symbols). The isotopic composition of recent global precipitation plots along the Meteoric Water Line (MWL). The Elk Point Basin Trend is the trend formed from the WD and δ^{18}O values of waters presently residing in formations in the Elk Point Basin. Dotted lines that parallel the Elk Point Basin Line represent calculated stable isotopic compositions of carnallite that would precipitate from basin fluids at 41°C and 25°C. A and B are the calculated isotopic compositions of fluids that most likely formed the carnallites in the ore zone.

SYLVITE-CARNALLITE RELATIONSHIPS

The occurrence of sylvite in ancient evaporites is problematic because chemical models do not predict the precipitation of sylvite from evaporating seawater (Harvie *et al.*, 1980), and direct sylvite precipitation from evaporating seawater is not observed in laboratory experiments (Brochert and Muir, 1964; Braitch, 1971). Because sylvite is a common mineral in saline deposits, alternate methods of primary sylvite precipitation have been proposed, such as the influence of meta-stable mineral equilibria, or strong depletion of the sulphate content of seawater by precipitation of sulphate minerals or biological reduction of sulphate in solution prior to chloride mineral precipitation (Holser, 1979 and references therein).

Carnallite may undergo incongruent dissolution to form sylvite by reaction with undersaturated fluids, a process that has been proposed to explain the occurrence of sylvite as a syn-depositional phase in the Prairie Evaporite and other evaporite deposits (see Lowenstein and Spencer, 1990, for review). Petrographic relations indicate that some sylvite and carnallite in the Prairie Evaporite have been reversibly transformed from one to the other and are younger than co-existing halite (Wardlaw, 1968; Fuzesy, 1983; Baadsgaard, 1987). Beds of sylvite generally overlie carnallitic regions, which is the reverse of an expected primary depositional sequence. In areas of predominantly sylvite, the overlying formations are observed to drape into the evaporite (Wardlaw, 1968), indicating that mineralogical changes responsible for stratigraphic thinning occurred after burial of the deposit. These petrographic relationships between sylvite and carnallite, along with stratigraphic evidence argue against a syn-depositional origin for most of the sylvite in this deposit, and favour a sub-surface diagenetic process.

Breakdown of carnallite to sylvite and a Mg-rich brine is favoured as the lithostatic pressure increases during burial of the deposit, although reaction of carnallites with undersaturated fluids in the subsurface can also result in sylvite and an Mg-rich brine (Koehler *et al.*, 1990). In any event, the breakdown of carnallite will result in the formation of highly concentrated, Mg-rich chloride brines.

The Middle Devonian Dawson Bay Formation, which unconformably overlies the Prairie Evaporite, has been affected by fluids that have interacted with the evaporite deposits, as demonstrated by halite-filled fractures and porosity (Dunn, 1982), as well as high bromine concentrations in interstitial brines (Wittrup and Kyser, 1990). It is possible that the pervasive dolomitization of Dawson Bay limestones may have been the result of interaction of Mg-rich brines derived from the breakdown of carnallite in the underlying Prairie Evaporite. While the source and character of dolomitizing fluids in basins have remained uncertain, it is evident that for massive replacement of limestones by dolomite, there must exist an external source of magnesium (Hardie, 1987). In this regard, brines resulting from the breakdown of carnallite to sylvite must be considered as potential dolomitizing fluids.

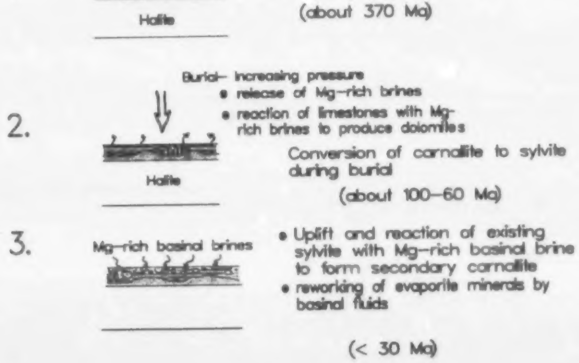
1.
Halite
Dominantly Halite–carnallite rock
(about 370 Ma)
Burial– increasing pressure
• release of Mg–rich brines
• reaction of limestones with Mg-rich brines to produce dolomites
2.
Halite
Conversion of carnallite to sylvite during burial
(about 100–60 Ma)
3.
Mg–rich basinal brines
• Uplift and reaction of existing sylvite with Mg–rich basinal brine to form secondary carnallite
• reworking of evaporite minerals by basinal fluids
(< 30 Ma)

Figure 4: Simple model of carnallite-sylvite relationships and fluids that have affected the Prairie Evaporite.

CONCLUSIONS

Stable isotopic compositions of carnallites indicate that carnallite in the Prairie Evaporite formed from, or equilibrated with, fluids which have their origins in formations above the evaporite, and do not represent primary precipitation from Middle Devonian seawater. At least three distinct fluid events can be discriminated from the hydrogen and oxygen isotopic composition of carnallite from the ore and anomalous zones in potash mines. These events probably represent continual reworking of the potash deposit by later fluids. It is likely that some of the carnallites formed by reaction of sylvite with Mg-rich brines.

Maximum burial of the deposit probably occurred during the Cretaceous, as reflected in fluid inclusion temperatures in halites from the same deposit (Chipley and Kyser, 1989). Radiometric ages of carnallites from the Alwinsal Willowbrook well and from the Patience Lake Member at PCS Lanigan indicate that some of the carnallite is the result of relatively recent processes (<30 Ma) (Baadsgaard, 1987).

Petrographic relations, experiments on evaporation of seawater, and stable isotopic compositions of the hydration water in carnallite, indicate that primary carnallite, formed in Middle Devonian time, may have dehydrated to form sylvite and a Mg-rich brine in response to increasing lithostatic pressure as the deposit was buried (Fig. 4). These resultant Mg-rich brines may have escaped the Prairie Evaporite to react with limestone sequences in the basin to form dolomites, such as in the Dawson Bay Formation. This reaction would significantly alter the primary porosity of these carbonates and their potential as hydrocarbon reservoirs.

ACKNOWLEDGEMENTS

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MINERALOGY AND GEOCHEMISTRY OF CRETACEOUS OIL SHALES IN MANITOBA

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ABSTRACT

Interest in organic-rich strata in the argillaceous Upper Cretaceous Niobrara and Favel Formations of the Williston Basin as a potential oil shale resource is indicated by leasing and test drilling in the outcrop area of western Manitoba. The rocks are typically dark brown to black, weakly indurated marlstones and claystones. In hand sample, they are structureless to thinly bedded with discontinuous, parallel stratification, and rare pyrite-filled burrows and fractures. Thin lenses of siltstone and bentonite are found locally. Thin section examination reveals very fine primary lamination. The sediments are moderately fossiliferous throughout and concentrations of foraminiferal and inoceramid debris are common. Inorganic components are mainly clay minerals, calcite, pyrite, and quartz. Clay minerals are dominantly illite and kaolinite with secondary abundances of various expandable lattice clays. Lesser amounts of gypsum and feldspar are also present but rarely make up more than 10 percent of the mineral suite.

The oil shales average about 6 percent total organic carbon with the richest sections giving oil yields of about 60 litres/tonne. The kerogen is immature to marginally mature as indicated by vitrinite reflectance and peak yield temperature data. Detailed petrographic examination and pyrolysis yield determinations suggest Type II (marine) and Type III (terrestrial) organic matter are intermixed, and show considerable stratigraphic and spatial variation in their ratios.

INTRODUCTION

The declining discovery rate of conventional oil and gas reserves in North America has prompted energy scientists to conduct research into the development of unconventional hydrocarbon fuel sources. In Canada much of this research is centred around large deposits of bituminous sand located in northeastern Alberta. In addition, coal, which preceded petroleum as a source of energy, is again becoming a prime candidate as a petroleum substitute in North America. However, another form of fossil organic matter known as "oil shale" is also being considered as a potential source of hydrocarbon fuel. The ability to extract oil from this type of rock has been known for almost three hundred years and estimates indicate the potential of oil shale as an energy source is comparable to that of coal (Montagna, 1978). Despite these facts, oil shale as a fossil fuel resource is little understood and, in North America, virtually unused.

"Oil shale" is an ambiguous term commonly used in an economic sense rather than in a lithologic or petrographic context. Oil shale usually refers to a rock's ability to yield oil upon heating and retorting. Thus, a variety of fine grained sedimentary rocks with or without interbedded shales are routinely classified as oil shales. The origin and depositional setting of these rocks are the subject of considerable speculation and have been discussed by many authors (*e.g.*, Cane, 1976; Duncan, 1976; Robinson, 1976; Montagna, 1978; Greensmith, 1978), most of whom agree that oil shales are generally formed as a result of deposition of inorganic silt and clay along with organic matter in conditions where: (a) organic matter is abundantly produced, (b) an anaerobic environment prevails, and (c) destructive organisms are absent. Depositional settings where these conditions are known to occur include: large stratified lake basins (Green River Formation), small bodies of water such as swamps, stagnant streams or lagoons (Australian Torbanite deposits), and shallow marine seas (Devonian/Mississippian black shales of eastern and central North America).

In Canada, the occurrence of various deposits of oil shale has been summarized in reports by Macauley (1981, 1984 a and b). One of the more significant and regionally extensive Canadian oil shale deposits is the organic-rich Upper Cretaceous shale of Manitoba and Saskatchewan. The oil shale crops out along the Manitoba Escarpment and extends westward in the subsurface into Saskatchewan and Alberta where it is a potential oil and gas source rock. These Cretaceous shales are economically attractive because of their great thickness and vast areal extent.

In recent years several companies have conducted preliminary economic and feasibility studies in Manitoba and Saskatchewan. Despite this interest, however, very little modern geochemical or petrographic analysis has been done to aid in assessing the shale's resource value. The purpose of this paper is to summarize the mineralogy, geochemistry, and stratigraphy on a regional scale.

PREVIOUS WORK

In Manitoba, the presence of shales and other fine grained rocks capable of yielding oil upon heating was first noted by Tyrrell (1892). Early attempts to quantify the richness of the Cretaceous oil shales in the region include studies by McInnes (1913), Clapp (1915), and Ells (1923). Lithologic descriptions of the shale in the outcrop area of Manitoba are provided by Kirk (1930) and Wickenden (1945). The speckled appearance of the shale has been studied by Goodman (1951), Langhus (1971), and Hattin (1975). Aspects of the biostratigraphy have been reported on by Park (1965) and North and Caldwell (1975). To date, the most definitive investigation of Manitoba's Upper Cretaceous sequence is that of McNeil and Caldwell (1981).

Preliminary assessments of the shales' mineralogy are provided by Bannatyne (1970) and Madden (1974). More recently, Hutton (1981a and b) has presented the only published petrographic examination of the organic matter of the shales from samples in Manitoba and Saskatchewan.

Much of what is known about the resource potential of the shale is derived from drilling and exploration programs carried out in the Manitoba Escarpment area during the 1960s. Most of this work involved estimating the hydrocarbon potential of the rock by Fisher assay analysis. Using these assay results, Beck (1974) estimated the recoverable oil from the shale for a portion of the Escarpment area to be about 400×10^6 m^3. This estimate was reduced to less than 200×10^6 m^3 by Macauley (1981). Further studies are currently being conducted using a dynamic retorting process which is designed to increase hydrocarbon yields (B. Ritzi, UMATAC Industrial Processes; pers.comm.).

GEOLOGIC SETTING AND STRATIGRAPHIC NOMENCLATURE

The study area encompasses the western part of Manitoba and portions of east-central Saskatchewan (Fig. 1). The eastern limit is defined by the Manitoba Escarpment which represents the erosional edge of Cretaceous strata which were deposited in the Western Interior Basin. The Escarpment trends northwesterly for 700 km and comprises five upland areas.

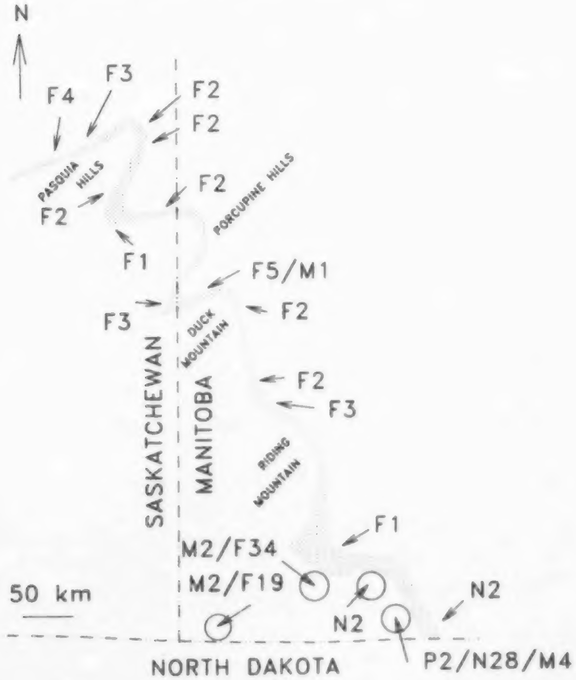

Figure 1: Location of samples used in this study. The shading shows the area of Upper Cretaceous oil shale outcrop/subcrop. The letters refer to the formation sampled (N=Niobrara; M=Morden; F=Favel) and the numbers refer to the number of samples analyzed from that location. The circles are core locations.

Sedimentation in the Western Interior Basin occurred within an epeiric sea which transected North America during the Cretaceous. Large transgresssive-regressive oscillations of the sea resulted in the widespread deposition of organic-rich sediments which today extend from the Manitoba Escarpment through the subsurface to the Cordillera region (Kauffman, 1977). Rice and Shurr (1980) recognized a series of north-south trending facies systems in the eastern part of the depositional basin. In Manitoba, the sediments are characteristically chalky and composed of fine grained carbonate material dominated by calcareous microfossils and nannofossils.

The stratigraphic nomenclature for the Upper Cretaceous of this part of North America varies considerably (*e.g.*, Kirk, 1930; Wickenden, 1945; Hutt, 1963; Gill and Cobban, 1965; Macauley, 1981; McNeil and Caldwell, 1981; Bannnatyne and Watson, 1982). That proposed by McNeil and Caldwell (1981) and summarized in Figure 2 is used in this paper. It differs from the nomenclature commonly used in Manitoba and eastern Saskatchewan in three important aspects: (i) Pierre Shale is used instead of the term Riding Mountain Formation; it also includes the Pembina Member of the Vermillion River Formation; (ii) Niobrara Formation is used in place of the Boyne Member of the Vermillion River Formation; and (iii) the Morden is elevated to formation status.

PIERRE	UNNAMED
	ODANAH
	MILLWOOD
	PEMBINA
	GAMMON FERR.
NIOBRARA	
MORDEN	
FAVEL	ASSINIBOINE
	KELD
ASHVILLE	

Figure 2: Lithostratigraphic nomenclature used in this paper.

The organic-rich shales in western Manitoba are in the Niobrara and Favel Formations. In the subsurface of Saskatchewan and Alberta, equivalent units are termed the First and Second White Specks, respectively. In Manitoba the intervening, non-calcareous, organic-poor Morden Shale separates the two organic-rich formations. Farther west, in north-central Saskatchewan, the Morden Shale is absent and the organic-rich oil shales form a single unit.

McNeil and Caldwell (1981) show that the thicknesses of the formations vary considerably from the southeast to northwest along the Escarpment. The overall northwestward thinning of the units is most likely due to either erosion or nondeposition.

MATERIALS AND METHODS

One hundred and sixty-seven samples taken from core and outcrop locations throughout western Manitoba and eastern Saskatchewan were used in this study (Fig. 1). Although all available cores of the appropriate units were examined and sampled, most subsurface stratigraphic information presented in this paper was derived from four long (130 m) cores taken in southern and western Manitoba. Collectively, these cores represent a complete section through the Favel, Morden, Niobrara, and Pierre Shale Formations. The outcrops sampled as part of this study were generally confined to the Favel.

All cores and outcrop sections were logged, photographed, and sampled. After description of colour, bedding, grain size, and macrofossil content, samples were analyzed for bulk mineralogy, detailed clay mineralogy of size fractionated material, and detailed carbonate mineralogy using X-ray diffraction techniques. Mineral identification was aided by the use of an automated search-match computer program (Marquart, 1986). The percentages of the various minerals in the bulk (whole rock) analyses were estimated from the diffractograms using a method modified after Schultz (1964; see Last, 1980). Replicate XRD analyses using this technique suggest that the precision of any individual mineral determination is about ±8 percent. Minerals which occur in trace amounts or small weight percents cannot generally be adequately identified using standard X-ray diffraction techniques (Last, 1980). Clay mineralogy of the finer than 2 micron size fraction was qualitatively and semi-quantitatively evaluated using techniques modified from Thorez (1975), Drever (1973), and Biscaye (1965). The inorganic fraction of the sediment was also examined by thin section petrography, scanning electron microscopy, and energy dispersive X-ray analysis. Our examination of the organic fraction of the sediment was designed to evaluate the regional and stratigraphic variation in: (a) organic matter richness, (b) organic matter type, and (c) organic maturity of the shale. Analytical techniques include thin section examination of isolated kerogen, vitrinite reflectance study, spore colouration, and pyrolysis yield determinations. Further details of sampling and analytical procedures, preparation and quantification techniques, and instrumentation can be found in Kovac (1985).

RESULTS AND DISCUSSION

INORGANIC PETROGRAPHY OF THE OIL SHALE UNITS

In general, the organic-rich rocks of the Cretaceous sequence in Manitoba are easily recognized by their dark grey to very dark brown colour, fissile nature, characteristic white speckled appearance, and distinct hydrocarbon odour from freshly broken rock. Locally, the organic-rich shales include thin bentonite seams, laminae and thin beds of inoceramid and oyster bivalve shells (Fig. 3), and/or light coloured, non-fissile, non-laminated zones of pyrite filled burrows. Lithologically, the oil shales vary between claystone and marlstone. The sediment is usually very finely laminated (Fig. 4), and beds range up to 0.05 m in thickness. Outcrops often appear massive on fresh exposures. Composition, organic content, and bedding characteristics of these rocks appear to have no consistent inter-relationship: shales with low organic matter content are as well laminated as those with high organic content, and organic-rich rocks range in texture from very finely laminated and fissile to massive. Discontinuous laminae and thin beds composed of inoceramid debris are also common.

Figure 3: Outcrop exposure of the Favel Formation at Ochre River, Riding Mountain area. Arrow points to a resistant limestone bed composed of shell fragments.

Figure 4: The Favel Formation showing fine light-dark lamination and the speckled appearance of the rock. The view is 4 cm across; the arrow points to a discontinuous light coloured lamina composed of inoceramid fragments and biogenic calcite prisms.

In thin section, the rock is finely laminated, even in samples which appear to be massive in hand specimen. The laminae comprise alternating layers of carbonate-rich sediment (light coloured) and organic-rich and clay-rich material (dark coloured). In all thin sections, the dark coloured matrix material is composed of mixtures of clay minerals, silt sized pyrite and quartz, and organic matter. The carbonate material is usually a mixture of three main constituents: (i) algal coccolithic debris, usually in the form of small (0.2 mm diameter) fecal pellets (the white specks), (ii) tests of foraminifera, and (iii) calcite fragments and prisms derived from inoceramid shells. The latter two components frequently show complete or partial replacment by pyrite.

Relative proportions of the three carbonate constituents vary greatly over short stratigraphic intervals. In some samples, the white carbonate laminae are dominated by foraminifera tests (Fig. 5) whereas in others they are composed entirely of calcite prisms. Relatively pure carbonate accumulations form thin limestone beds within the Favel Formation. They are made up of the same three constituents present in light-coloured, carbonate-rich shale laminae.

The origin of the white specks which are so common in the sequence has generated an unusual amount of discussion in the literature (*e.g.*, Goodman, 1951; Langhus, 1971; Hattin, 1975). Our thin section and SEM examination indicates that the white specks are composed nearly entirely of closely packed coccoliths and coccolithic fragments (Fig. 6). Foraminiferal debris is rare. We agree with Hattin (1975) that the former are most likely fecal pellets from small herbivorous organisms.

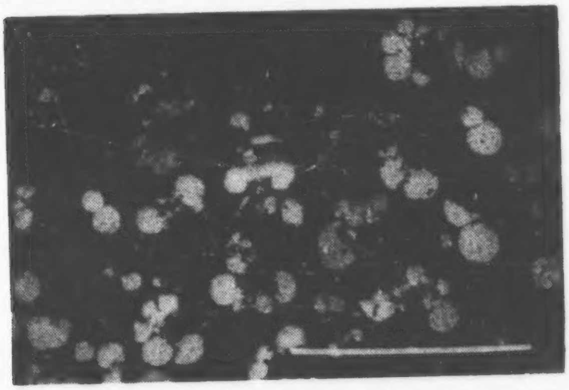

Figure 5: Photomicrograph showing a carbonate-rich lamina composed almost entirely of foraminifera tests. Sample is from the Favel Formation at 482.5 m depth in core 11-29-1-25WPM; scale bar is 0.5 mm.

The shales are composed of a relatively simple suite of minerals: clay minerals, calcite, quartz, pyrite, gypsum, and feldspars (Table 1). A similar bulk mineral suite was identified by Bannatyne (1970). Our mineralogical results are in striking contrast with the results of analyses for these shales provided by Macauley (1984a) and Macauley *et al.* (1985), who report a much greater variety of minerals in this Upper Cretaceous sequence. Presumably most of the nearly 50 "secondary minerals" listed by Macauley (1984a) occur in very small or trace amounts, and are therefore not identifiable using our standard X-ray diffraction analysis techniques.

The clay mineral fraction of the oil shales consists of kaolinite, illite, smectite, and mixed layer clays. The kaolinite and illite groups are dominant, with only secondary proportions of smectite and mixed layer clays (Table 1). The average clay mineral values are somewhat misleading because they include mineralogical analyses of bentonite beds, which are made up of nearly 100 percent smectite. Eliminating the few bentonite bed samples gives the following average clay mineral percentages for the oil shales: 51.5 kaolinite, 38.5 illite, 5.5 mixed layer, and 4.5 smectite. A similar suite of minerals was identifed by Madden (1974) for the Favel, Morden, and Niobrara formations. Again, however, our clay mineralogy results differ from those of Macauley (*op. cit.*) who reports that these shales are dominated by montmorillonite with little or no kaolinite present. In our analyses, only six of the nearly 150 samples contained less than 25 percent kaolinite in the clay sized fraction.

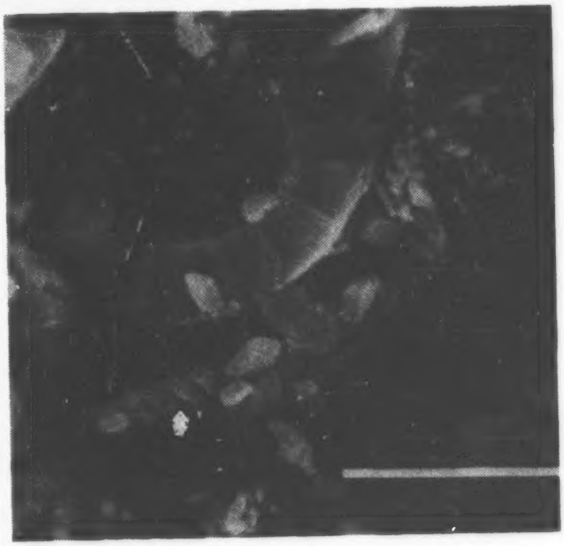

Figure 6: Scanning electron micrograph showing the coccolithic debris found within a white speck. Sample is from the Favel Formation at 75.3 m depth in core 16-11-8-11WPM; scale bar is 15 microns.

Table 1: Average mineralogical composition of the oil shales (values in percent; mean whole rock values are based on 108 samples analyzed; mean clay mineralogy values are based on 42 samples).

	Favel Formation	Niobrara Formation
Whole Rock Mineralogy		
Clay Minerals	40.3	52.7
Calcite	37.0	22.2
Quartz	7.2	10.3
Pyrite	9.6	8.5
Gypsum	3.8	3.1
Feldspars	2.1	3.1
Clay Mineralogy		
Kaolinite	37.9	40.2
Illite	32.8	35.7
Smectite	22.6	12.9
Mixed Layer	6.8	11.2

The character and composition of the sediments are important factors in attempting to evaluate the resource potential of these oil shales. The mineral constituents of the shales are conveniently subdivided into three groups on the basis of origin: (i) authigenic minerals including pyrite and gypsum, (ii) biogenic minerals, predominantly carbonates derived from coccolithic fecal pellets, foraminifera tests, and shell fragments, and (iii) detrital minerals such as clay, quartz, and feldspars. The origin of the kaolinite in the samples is not completely understood. Pollastro (1981) suggested that the kaolinite in the Niobrara Formation in Colorado is authigenic (specifically, an *in situ* alteration product of volcanic ash). Our SEM examination revealed that, in Manitoba, textural and fabric relationships suggest at least some of the kaolinite may be authigenic in origin.

By treating each of these mineral components as end-members, the stratigraphic variability in the oil shales has been evaluated using a triangular diagram. The various stratigraphic units contain similar overall abundances of these components (Fig. 7), although the Favel shales tend to have a higher proportion of biogenic minerals and the Morden Formation samples are dominated by detrital components.

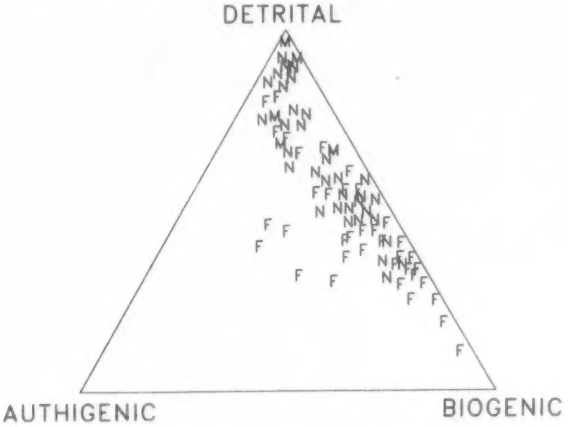

Figure 7: Trilinear plot of the whole rock mineralogy of the oil shales. The letters correspond to the formation name (F=Favel, M=Morden, N=Niobrara).

The stratigraphic variation in mineralogy within each oil shale sequence is pronounced in both cores and outcrops. Figure 8 shows the variation in the three mineral groups through the Favel Formation in one of the cores from western Manitoba. The bottom and top of the Favel are dominated by detrital minerals whereas the middle of the formation is mainly made up of biogenic carbonates. In contrast, little systematic variation in the relative abundance of the various components within the clay mineral fraction of the Favel is found in this core. With the exception of three samples collected over a metre, the illite and kaolinite groups constitute 85 to 90 percent of the total clay mineral suite.

ORGANIC GEOCHEMISTRY AND PETROGRAPHY OF THE OIL SHALE UNITS

ORGANIC MATTER RICHNESS

The total organic carbon (TOC) for samples from the two oil shale units ranges from about 1 percent to just over 11 percent with an overall average value of approximately 6 percent. In cores, TOC correlates well with the degree of bioturbation in the sediments: beds which are moderately to well bioturbated have low organic carbon contents. This correlation does not persist into the outcrop areas, where samples devoid of white specks and obvious lamination may have high TOC values.

The richness of the oil shales has also been evaluated by bulk sample pyrolysis ("Rock-Eval"). This technique quantifies the amounts of hydrocarbon which are: a) present in the rock through "natural" maturation (the relatively low temperature S-1 peak), and b) produced during thermal conversion of the kerogen (the higher temperature S-2 peak). The sum of these volumes is referred to as the petroleum potential of the rock and is normally expressed as weight of hydrocarbon per weight of rock (*i.e.*, kilograms hydrocarbon per tonne of rock).

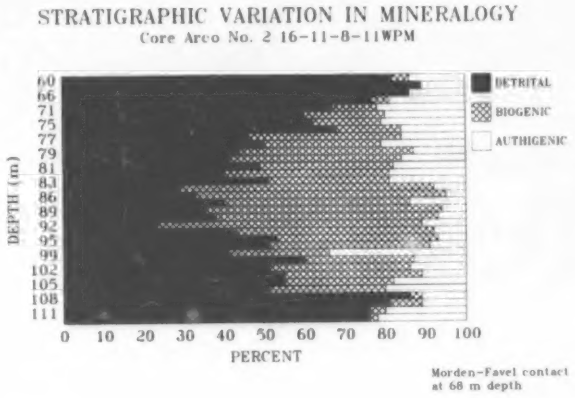

Figure 8: Stratigraphic variation in mineralogy.

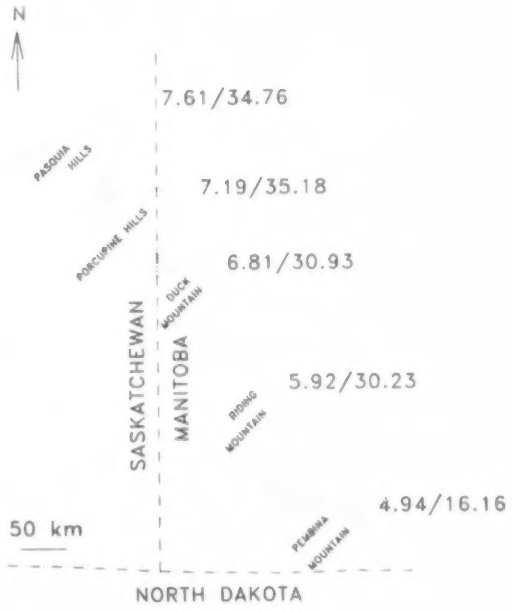

Figure 9: Spatial variation in richness. The first number of the pair is the average TOC value (percent) for the shales from the adjacent Escarpment area; the second number of the pair is the average petroleum potential (kg/tonne) for the shales from that area.

Petroleum potentials for the organic-rich beds range from a minimum of 1.68 kg/tonne to a maximum of 51.6 kg/tonne, with an overall mean of about 30 kg/tonne. These values convert to approximately 1.7 litres/tonne, 53.7 litres/tonne, and 30.2 litres/tonne, respectively.

The organic richness of the shales also shows great variation over short stratigraphic distances. The Favel Formation is slightly richer than the Niobrara Formation. Figure 9 shows the spatial variation in average richness along the Manitoba Escarpment, revealing a distinct increase from southeast to northwest.

ORGANIC MATTER TYPE

In normal, reflected light microscropy of thin sections, organic matter in the shale appears as dark brown to black lenses and wisps up to 0.04 mm thick. Examination of the isolated kerogen shows that the organic matter consists mainly of amber to yellow spherulitic amorphous material with secondary abundances of grey amorphous material and inertinite (Table 2; classification and terminology after Masran and Pocock, 1981). The amorphous nature of the organic matter prohibits identification of either a terrestrial or marine source.

Table 2: Organic petrography of the oil shales (values in percent of total isolated kerogen; based on six samples analyzed).

Organic Matter Type	Favel Formation	Niobrara Formation
Structured/terrestrial	1	1
Pollen/Spores	0	1
Inertinite	5	2
Biodegraded/terrestrial	0	0
Amorphous/yellow/amber	65	64
Amorphous/grey	29	32
Structured/marine	0	1

Organic matter type can also be approximated using data from pyrolysis yield determinations. As shown in Figure 10, the samples generally fall between organic matter Type II (marine) and Type III (terrestrial).

Stratigraphic variation in organic matter type is inconsistent, although samples from the Favel Formation tend to be closer to the Type II line, those from the Niobrara Formation closer to the Type III kerogen line. Spatially, Type II kerogen content is generally increasingly dominant northwesterly along the Escarpment.

ORGANIC MATURITY

Peak yield temperatures (T_{max} or the temperature of maximum pyrolysis yield) range from 400°C to 430°C. These values are below the thermally mature range for petroleum source rocks (435°C to 460°C). Vitrinite reflectance studies, however, show a range from 0.41 percent R_O to 0.54 percent R_O, which indicate marginally mature sediment (assuming Type II organic matter; Tissot *et al.*, 1980). Similarly, spore coloration ranges from yellow to yellow-orange, suggesting an immature to marginally mature rank. No consistent stratigraphic or spatial trends in these maturity indicators have been recognized.

SUMMARY

1. The Upper Cretaceous oil shales in the Manitoba Escarpment area are grey to brown, moderately indurated, claystone to marlstone.
2. Most of the organic rich beds exhibit a white speckled appearance. The white specks are composed of coccoliths and coccolithic fragments and most likely are fecal pellets.
3. The oil shales are usually finely laminated with alternating light coloured fecal pellets, foraminifera tests, and shell fragments and dark coloured clay minerals, silt sized pyrite and quartz, and organic matter.
4. The shales are a relatively simple mixture of clay minerals, calcite, pyrite, quartz, gypsum, and feldspars. The proportions of the two major constituents (clay minerals and calcite) show an inverse relationship. Although the Favel Formation is slightly more calcareous than the Niobrara oil shales, stratigraphic variation in mineral abundance is generally inconsistent.

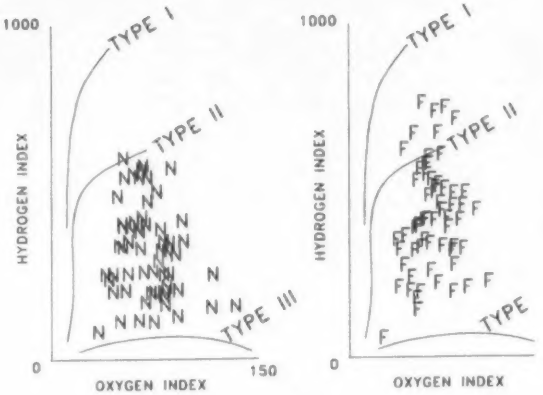

Figure 10: Organic matter typing by Rock-Eval pyrolysis. N refers to samples of Niobrara Formation; F refers to samples of Favel Formation. The oxygen index values are in units of mg CO_2/g organic carbon; the hydrogen index values are in units of mg total hydrocarbon/g organic carbon.

5. The clay mineral fraction of the shales is composed mainly of kaolinite and illite, with smaller proportions of smectite and mixed layer clays.
6. TOC values average about 6 percent and petroleum potentials about 30 kg/tonne. Favel Formation samples are slightly richer than those from the Niobrara Formation. Richness increases slightly northwestward along the Escarpment.
7. Petrographic examination of isolated kerogen indicates that nearly all of the organic matter is amorphous. Pyrolysis yield determinations suggest a mixture of Type II and Type III organic matter. Stratigraphic trends in organic matter type are inconsistent apart from a slightly higher content of Type II kerogen in oil shales at the northwestern end of the Escarpment than in those at the southeastern end.
8. Vitrinite reflectance and spore coloration determinations indicate that the oil shales in the Manitoba Escarpment area are immature to marginally mature. Peak yield temperatures from pyrolysis suggest immaturity on the assumption of Type II organic matter.

ACKNOWLEDGEMENTS

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SUBSURFACE BRINES IN SOUTHERN SASKATCHEWAN

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ABSTRACT

Subsurface brines with anomalously high calcium and magnesium content are widespread in reservoirs of Cambrian to Devonian age in southern Saskatchewan. Geographically, the anomalous brines are reported most commonly in the potash mining area around Saskatoon (lat. 52° N, long. 106° 30' W) and in the southeastern part of the province (lat. 49° 30' N, long. 103° W). Some of the brines occur as potash-mine inflow and these have been shown to have anomalously high contents of bromine, iodine and strontium. Origin of the brines is uncertain but they appear to have undergone a complex evolutionary history that may have involved dissolution of evaporite minerals in the Prairie Formation.

Data on characteristics of the reservoirs containing the brines are limited. Analysis of mechanical logs and a few drillstem tests shows that some of the reservoirs have significant thickness and permeability, and at least one may be large enough for commercial development.

In the Michigan Basin subsurface brines provide the raw material for a billion dollar chemical industry. Preliminary work indicates that some of the Saskatchewan brines have calcium, magnesium, bromine and iodine contents that compare favourably with concentrations of these components in the commercial brines of the Michigan Basin.

INTRODUCTION

Preliminary investigations by Kreis and Gent (in press) suggested that subsurface brines of unusually high quality tend to be associated with Middle Devonian and older formations of southern Saskatchewan. A paucity of water analyses data in these formations made it necessary to combine stratigraphic units on the basis of their physical, chemical and hydrogeological characteristics. Four such units were recognized: the Manitoba Group (Souris River and Dawson Bay formations), Elk Point Group (Prairie, Winnipegosis and Ashern formations), the combined strata of the Interlake to Red River formations, and the Winnipeg Formation to the regolith on the Precambrian basement (Fig. 1).

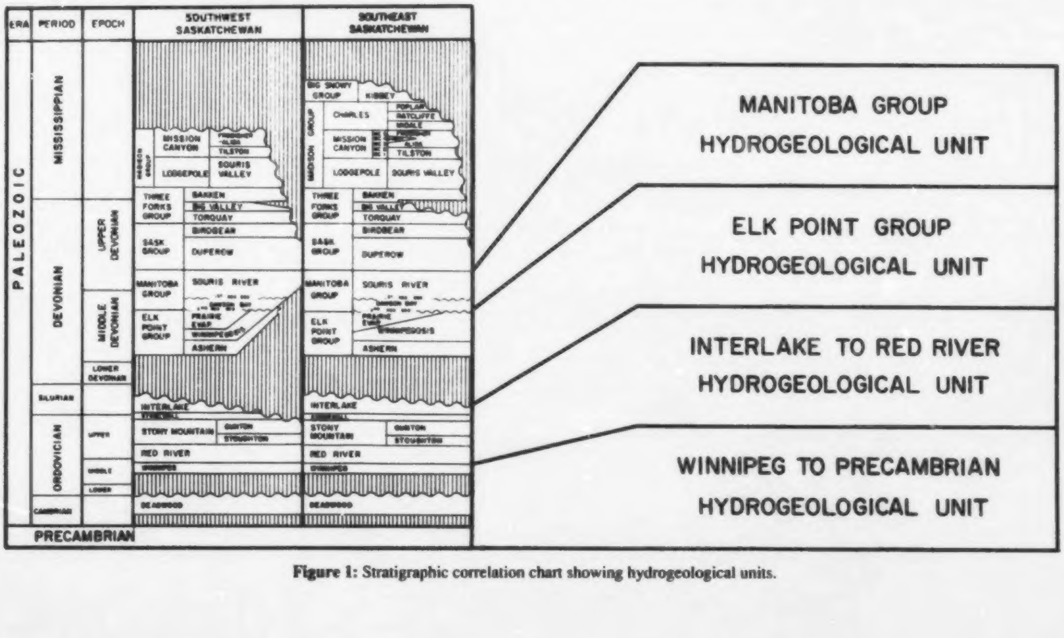
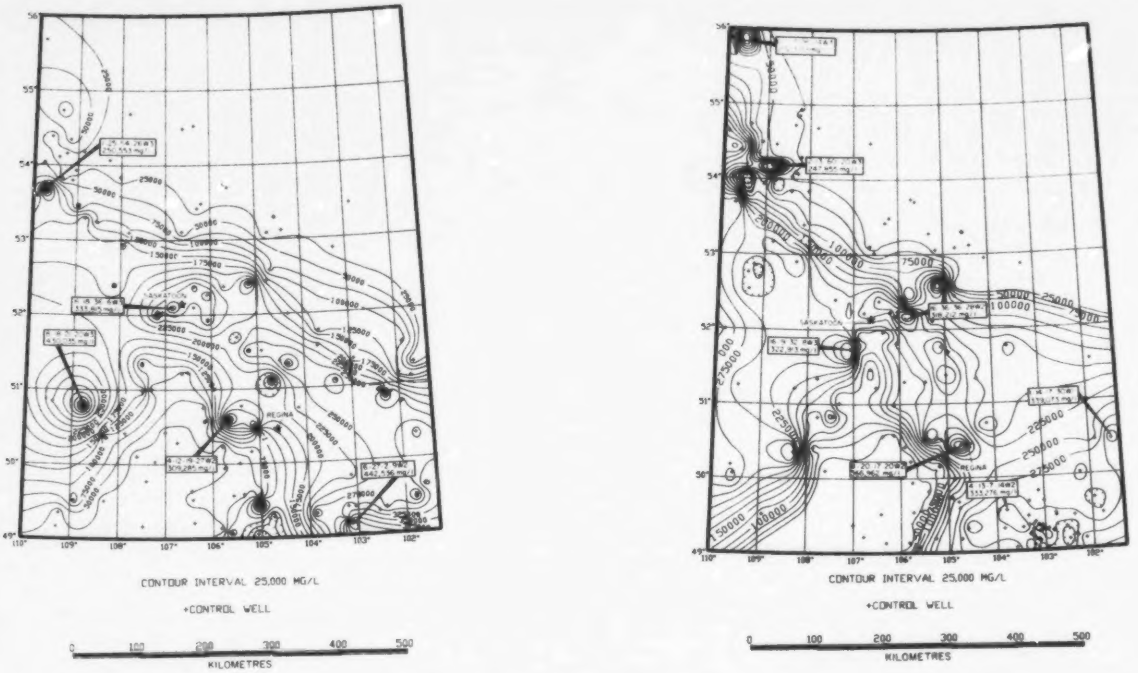
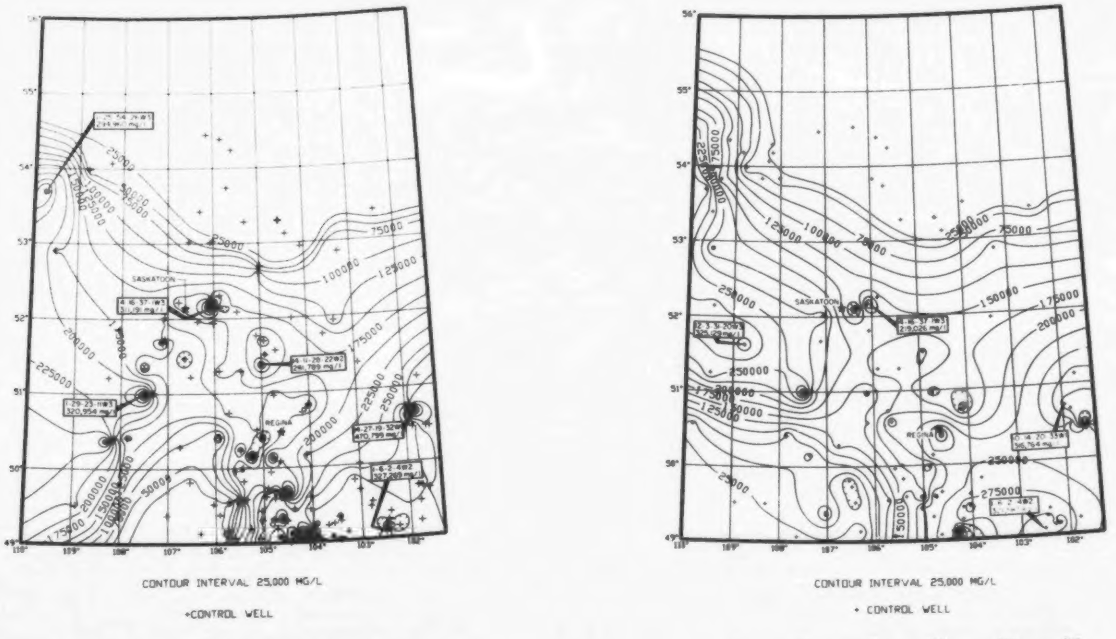


Figure 1: Stratigraphic correlation chart showing hydrogeological units.



SOURIS RIVER/DAWSON BAY
TOTAL DISSOLVED SOLIDS

ASHERN/MEADOW LAKE BEDS
WINNIPEGOSIS/PRAIRIE EVAP
TOTAL DISSOLVED SOLIDS



INTERLAKE/STONEWALL
STONY MOUNTAIN/RED RIVER
TOTAL DISSOLVED SOLIDS

PRECAMBRIAN/EARLIE/DEADWOOD
WINNIPEG TOTAL DISSOLVED SOLIDS

Figure 2: Maps showing distribution of total dissolved solids in all four hydrogeological units.

In this paper, the term "brine" refers to saline formation waters in the deep sedimentary basin. The basic water chemistry data used for mapping were derived from water chemical analyses stored in the Well File data base at Saskatchewan Energy and Mines. These data are now available in digitized form. Most analyses are on water samples from drillstem tests (DSTs) and, to a lesser extent, from perforated intervals. A few samples came from sources such as oil treater facilities or directly from the wellhead. The water chemistry data for potash mine inflows shown in Table 1a came from papers by Ogu and Arnold (1989) and Wittrup and Kyser (1990).

Sodium, potassium, calcium and magnesium are the four major cations in typical subsurface water. Calcium and magnesium ion concentrations are derived by laboratory analysis. Prior to 1980, potassium and sodium were commonly grouped together and values were calculated from the total ion balance. Recently, increased usage of KCl muds has caused laboratories to analyse for Na and K separately. Chloride, sulphate, bicarbonate and carbonate are the four major anions in a typical water analysis.

Iodide, bromide and boron ion concentrations are used in "hydrocarbon proximity" studies whereas barium and strontium are used in "scaling tendency" studies (Johnson, 1991). Unfortunately, these less abundant ions are rarely reported in the Petroleum Well File data base and their distribution is unknown.

ORIGIN OF SUBSURFACE BRINES

A discussion of the origin of brines in this paper is based on a compilation of relevant literature. Generally, salinities of formation waters increase with depth in a sedimentary basin (White, 1965; Dickey, 1969; Dinghman and Angino, 1969; Graf, 1982). A number of processes have been invoked to explain this increase. Early work suggested that membrane filtration by fine-grained sediments was an important factor (White ,1965; Berry, 1969; Hitchon *et al*., 1971). However, recent work on the stable isotopic composition of fluid inclusions in evaporite minerals by Knauth and Beeunas (1986) indicated a process whereby subsurface formation waters may have evolved by a mixing of

Table 1a: Chemistry of brines from selected mine-inflows.

Sample Location	Depth	Ion Concentration in Milligrams per Litre (mg/l)												Sp. Gr.
	(m)	Cl	Na	K	Ca	Mg	HCO3	Br	I	F	Li	Sr	pH	(gm/cm3)
Cory Mine														
Duperow Fm. (Shaft)	627	29,600	17,000	430	2,202	670	n.a.	12,900	12	0.100	n.a.	n.a.	6.5	n.a.
Souris River Fm. (Shaft)	847	101,000	42,000	940	24,500	4,800	1,464	863	127	0.050	n.a.	n.a.	8.2	1.112
Souris River Fm. (Shaft)	866	160,000	103,500	2,220	21,740	5,650	104	1,720	101	0.005	n.a.	n.a.	6.9	1.155
Souris River Fm. (Shaft)	890	171,000	81,000	1,890	25,300	4,750	207	2,160	114	0.004	n.a.	n.a.	7.2	1.185
Souris River Fm. (Shaft)	900	199,000	130,000	2,290	22,640	1,500	183	2,800	108	0.030	n.a.	n.a.	7.6	1.199
Souris River Fm. (Shaft)	900	194,000	173,000	1,510	18,950	4,770	244	2,460	118	0.003	n.a.	n.a.	7.5	1.193
Souris River Fm. (Shaft)	930	207,000	51,000	7,100	50,720	3,700	146	2,200	273	0.002	n.a.	n.a.	6.0	1.217
Souris River Fm. (Shaft)	930	228,000	60,000	6,300	68,500	18,300	110	5,950	330	0.002	n.a.	n.a.	5.6	1.240
Dawson Bay Fm. (Shaft)	975	355,000	54,870	16,300	18,200	28,400	n.a.	21,700	508	0.002	n.a.	n.a.	5.7	n.a
mine-level B-324, entry 3200		320,000	7,770	14,400	170,200	30,800	0	22,400	457	0.002	n.a.	n.a.	4.4	1.392
mine-level B-324, entry 3200		342,000	3,710	12,400	152,400	24,500	158	20,200	508	0.002	n.a.	n.a.	4.3	1.354
mine-level B-407, entry 4300		326,000	6,370	14,300	169,900	25,700	232	16,400	508	0.002	n.a.	n.a.	4.4	1.347
mine-level B-407, entry 4300		340,000	3,250	11,800	151,500	25,800	0	21,600	603	0.002	n.a.	n.a.	4.5	1.354
mine-level B-694, entry 6200		346,500	5,000	20,000	131,000	12,700	n.a.	17,800	n.a.	n.a.	n.a.	n.a.	4.4	1.343
mine-level B-695, entry 6100		346,900	7,060	25,800	149,000	36,000	n.a.	17,500	n.a.	n.a.	n.a.	n.a.	4.4	1.342
mine-level B-699, entry 6000		362,400	4,540	20,500	141,000	10,400	n.a.	17,800	n.a.	n.a.	n.a.	n.a.	4.3	1.351
mine-level P-6-12X, entry 6200		308,000	3,350	11,900	165,600	19,900	290	21,100	121	0.002	n.a.	n.a.	4.5	1.349
mine-level P-6-12X, entry 3200		319,000	4,020	13,300	160,700	20,500	122	19,700	400	0.002	n.a.	n.a.	4.5	1.341
mine-level P-6-12X, entry 6300		342,000	4,120	18,500	147,000	9,400	n.a.	18,800	n.a.	n.a.	n.a.	n.a.	4.4	1.348
mine-level entry 200		363,000	1,680	10,000	134,700	19,600	165	19,800	248	0.002	n.a.	n.a.	4.7	1.342
PCS-C613		328,000	5,820	19,400	139,000	18,000	n.a.	6,800	47	n.a.	17.00	2300	n.a.	1.365
PCS-L454		279,000	9,870	18,700	10,300	76,400	7	2,300	62	n.a.	1.10	110	n.a.	1.276
PCS-L772		279,000	9,570	18,300	14,400	75,600	18	2,600	88	n.a.	0.93	240	n.a.	1.283
Lanigan Mine														
mine-level Branch 6		271,000	11,000	12,300	8,470	88,000	0	8,630	46	0.002	n.a.	n.a.	3.4	1.256
mine-level Branch 7, S.E. Sump		273,000	10,200	11,200	4,500	83,900	0	8,800	83	0.002	n.a.	n.a.	5.1	1.264
mine-level Branch 7		277,000	11,980	13,400	5,340	84,900	73	9,320	67	0.002	n.a.	n.a.	5.3	1.250
mine-level bin area under Dawson Bay Fm.		367,000	3,200	5,100	180,600	17,800	0	28,500	806	0.002	n.a.	n.a.	4.0	n.a.
mine-level bin area under salt back		325,000	2,750	7,100	141,000	782	0	20,500	782	0.002	n.a.	n.a.	4.4	1.286
mine-level drill hole, Jan. 1987		290,300	9,800	25,800	149,000	93,300	n.a.	5,600	n.a.	n.a.	n.a.	n.a.	4.5	n.a.
Allen Mine														
Souris River Fm.	841	136,000	73,000	2,400	2,370	2,400	n.a.	1,170	192	n.a.	n.a.	n.a.	6.7	1.178
mine-level June 10, 1986		303,650	8,510	41,000	33,600	78,200	n.a.	2,640	n.a.	0.002	n.a.	n.a.	4.3	1.284
mine-level Sept., 1986		282,400	15,300	21,600	40,300	148,000	n.a.	5,500	n.a.	n.a.	n.a.	n.a.	4.3	1.275
Rocanville Mine														
Dawson Bay Fm.	890	199,000	143,000	2,160	7,620	2,000	48	2,200	101	0.003	n.a.	n.a.	7.1	1.196
mine-level panel 408, room 10		187,000	57,000	32,300	6,080	21,967	n.a.	5,000	77	0.002	n.a.	n.a.	5.9	1.230
mine-level panel 408, room 7		229,000	109,000	59,700	9,940	6,100	152	4,230	92	0.003	n.a.	n.a.	5.8	n.a.
mine-level panel 405, extension 0023		222,000	105,200	57,700	9,570	2,900	37	500	71	0.004	n.a.	n.a.	6.0	1.241
mine-level 408-00-07, 1986		229,000	74,400	65,000	7,210	5,790	n.a.	4,400	n.a.	n.a.	n.a.	n.a.	4.7	1.228
mine-level 408-00-01, 1986		238,200	45,900	50,000	5,930	19,700	n.a.	4,700	n.a.	n.a.	n.a.	n.a.	4.4	1.236
mine-level 405-00-23, 1986		242,200	55,400	53,200	18,900	5,700	n.a.	6,600	n.a.	n.a.	n.a.	n.a.	4.3	1.239

meteoric waters and ancient evaporite brines. This could explain the gradual increase in total dissolved solids (TDS) with depth, especially in larger evaporitic basins, but not the usually high (Ca+Mg)/Na ratio in calcium- and magnesium-rich brines. Graf (1982) suggested that formation fluids which have undergone membrane filtration show greatly increased (Ca+Mg)/Na ratios. Wittrup and Kyser (1990), however, argued that formation fluids are not strongly confined in the Elk Point Basin of Saskatchewan and the pressures required for membrane filtration are unlikely to have existed.

Using stable isotope work on floodwaters in potash mines in Saskatchewan, Wittrup and Kyser (1990) described three different types of brines and their origins: "(1) halite- and sylvite-saturated basinal brines from Devonian formations directly above the deposits, which are predominantly connate waters; (2) undersaturated waters of predominantly meteoric origin, from stratigraphically higher aquifers such as the Cretaceous Mannville Group; and (3) Ca-rich brines that may represent Devonian fluids associated with recrystallization of the evaporites."

It is noteworthy that Wittrup and Kyser (1990) found that the Saskatchewan potash mine inflows contained a significant connate oxygen content and that different mines exhibited different proportions of solutes in these waters. These relationships suggest that the high mineral content of many of the potash mine brine inflows is not the result of dissolution of salts along the mine workings alone, but likely includes significant amounts of connate waters as well.

Isotopic work done by Koehler *et al.* (1990) indicated an intrinsic relationship between evaporites in the Prairie Formation and waters in overlying formations. Petrological evidence and the distribution of bromine in the Prairie Evaporite minerals support the concept that the evaporite minerals have undergone repeated recrystallization in response to passage of hypersaline fluids through the formation (Schwerdtner, 1964; Wardlaw and Watson, 1966). The principal reaction appears to be the conversion of carnallite to sylvite (Holter, 1969; Koehler *et al.*, 1990):

$$KMgCl_3.6H_20 = KC1 + Mg^{2+} + 2C1^- + 6H_20.$$
$$\text{Carnallite} = \text{Sylvite} + MgCl_2\text{-rich solution}$$

The magnesium-rich solution produced by this reaction is believed by some authors (*e.g.*, Buchinski, 1988) to have reacted with the overlying carbonates of the Dawson Bay Formation to form dolomite and Ca-rich brine. Fuzesy (1983) argued that sufficient magnesium was not available from the conversion of carnallite to produce the magnesium chloride-rich brines found in the Prairie Evaporite. He suggested that large quantities of magnesium-rich dolomitizing fluids from the Winnipegosis Formation and fluid reactions with clay minerals contributed significant amounts of the magnesium.

Hardie (1990) described a process whereby calcium chloride brines are produced along zones of upwelling in modern active rift and transtensional basins. Some authors believe that many or all cratonic basins are underlain by ancient rift systems (Bally, 1980; Bally and Snelson, 1980). There is a probability that some of the evaporite deposits and brine solutions in the cratonic Williston Basin were derived from hydrothermal calcium chloride-rich brines originating from the Precambrian basement.

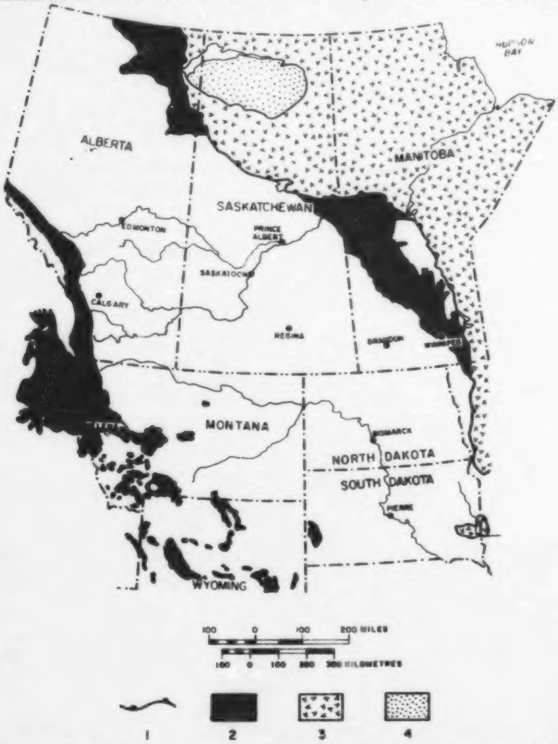


Figure 3: Major areas of recharge and discharge for formation waters in the Williston basin (after Simpson and Dennison, 1975).

DISTRIBUTION OF BRINES

Total dissolved solids (TDS) maps of the four hydrogeological units defined in this paper show a number of striking similarities (Fig. 2). For all units there are less than 25,000 mg/l of total dissolved solids in the northern third of the map area, where these units crop out or are covered by a few hundred metres of Lower Cretaceous strata and Pleistocene sediments (Simpson and Dennison, 1975). This region of low salinity formation waters reflects meteoric intake.

Another region of low salinity for all hydrological units is in southwestern Saskatchewan between longitudes 106°W and 107°W and latitudes 49°N and 50°N (Fig. 2). Porter and Fuller (1959), Hitchon (1964), and Simpson and Dennison (1975) have previously identified this region in Middle Devonian and older strata. The region is also coincident with a high fluid potential field for the Winnipeg and Winnipegosis formations (Hannon,

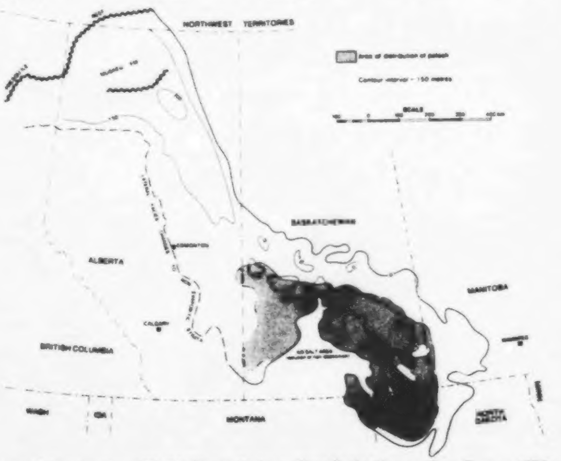


Figure 4: Distribution of Prairie Evaporite in southern Saskatchewan (from Fuzesy, 1982).

Table 1b: Chemistry of brines from selected boreholes.

Sample Location					Formation	Depth (m)	Ion Concentration in Milligrams per Litre (mg/l)										
Lsd	Sec	Twp	Rge	M			TDS	Cl	Na	K	Na+K	Ca	Mg	CO3+HCO3	SO4	pH	Sp. Gr. (gm/cm3)
6	27	2	9	2	Dawson Bay	2,422.0 - 2,435.0	442,536	287,000	11,000	13,060	-	119,300	11,420	537	214	5.9	1.320
8	18	21	20	3	Souris River	1,574.3 - 1,586.5	430,030	274,800	50,009	n.a	-	93,053	11,470	520	178	5.1	1.250
4	18	35	8	3	Souris River	969.3 - 993.6	327,636	207,000	32,000	11,300	-	64,500	11,900	936	0	8.4	n.a.
6	18	36	6	3	Souris River/Dawson Bay	899.8 - 985.4	333,815	215,700	46,300	4,481	-	53,000	13,700	447	187	6.9	1.260
7	24	4	5	2	Souris River	2,084.9 - 2,097.7	301,901	187,760	72,234	n.a	-	37,290	3,770	267	580	6.3	1.233
2	4	4	11	2	Stonewall/Stony Mountain	2,762.7 - 2,801.1	312,441	194,330	n.a.	n.a	78,100	36,650	3,160	30	170	5.4	1.226
6	13	2	19	2	Stony Mountain	2,953.5 - 2,962.7	315,228	195,000	n.a.	n.a	82,552	33,802	3,060	170	644	5.5	1.223
13	24	2	3	2	Deadwood	2,956.0 - 2,974.9	286,771	176,486	n.a.	n.a	75,968	33,759	n.a.	146	412	5.7	1.209
9	34	3	4	2	Deadwood	2,857.2 - 2,869.7	301,672	186,670	n.a.	n.a	81,516	30,940	2,220	66	260	5.9	1.213
8	20	17	20	2	Winnipegosis	uncertain	366,962	252,395	n.a.	n.a	25,732	36,161	51,303	1,293	77	5.2	1.264
4	16	18	23	2	Dawson Bay	1,524.0 - 1,542.9	259,930	169,400	62,900	n.a	-	3,480	22,820	480	850	5.5	1.224
4	1	4	7	2	Winnipegosis	2,458.0 - 2,471.0	315,857	194,083	101,600	6,090	-	4,400	8,564	304	815	6.7	1.211
3	20	17	30	1	Stony Mountain/Red River	1,224.1 - 1,297.2	347,733	213,257	n.a.	n.a	121,564	2,475	7,977	73	2,387	6.9	1.231
8A	22	2	9	2	Winnipegosis	2,581.0 - 2,584.0	249,496	141,000	22,810	66,320	-	11,210	6,682	488	981	4.5	1.175
1	14	17	30	1	Winnipegosis/Ashern/Interlake	1,076.0 - 1,181.1	339,073	206,976	n.a.	n.a	120,522	1,857	6,851	83	2,783	6.8	1.234
6	28	3	12	2	Red River	2,849.0 - 2,865.0	323,241	200,000	86,500	3,500	-	28,200	4,740	n.a.	301	5.5	1.212
13	26	20	33	1	Dawson Bay	870.0 - 891.0	253,689	145,776	58,140	40,544	-	1,875	4,532	n.a.	2,831	8.4	1.198
12	16	36	3	3	Dawson Bay	972.0 - 995.0	257,836	144,448	72,261	35,200	-	1,892	180	91	3,764	8.1	1.164

1987). Regional flow of subsurface formation waters in these and other Paleozoic formations in southern Saskatchewan is dominantly northeastward (Simpson and Dennison, 1975; Hannon, 1987). The Black Hills of South Dakota, the northern and eastern flanks of the Big Horn Mountains of Wyoming, the Little Belt, Big Snowy, and Judith Mountains of central Montana; and the Little Rockies of northeastern Montana have been considered to be major recharge areas for formation waters in the Williston Basin (Hannon, 1987) (see Fig. 3). Hitchon (1969) also considered the Rocky Mountain Foothills and the Cypress Hills to be important recharge areas. Discharge areas for these formation waters are located in east-central Saskatchewan and along the western shore of Lake Winnipegosis in southern Manitoba.

The region of low salinity (<25,000 mg/l) on all TDS maps in southwestern Saskatchewan coincides with an area from where the evaporite deposits of the Prairie Formation are missing. This area of 'No Salt' is interpreted to be the result of subsurface dissolution of salt (Holter, 1969; Fuzesy, 1982) by recharge waters.

In contrast, regions of high TDS (*i.e.*, >150,000 mg/l) in all four hydrogeological units shown in Figure 2 closely match the the Prairie Evaporite body (Fig. 4). The coincidence of the high TDS distribution in all four stratigraphic units suggests that there is a hydrological continuity between them.

BRINE ANOMALIES

Brines with 20,000 mg/l of calcium and 3,000 mg/l of magnesium are considered to be minimal for regional exploration (Hitchon and Holter, 1971) (Table 2). These concentrations represent approximately one quarter of the value for calcium (75,000 mg/l) and one third that of magnesium (10,000 mg/l) reported from a brine used in a commercial brine recovery operation of the Dow Chemical Company at Midland, Michigan (Table 3).

Local areas of anomalous concentrations of calcium and magnesium shown on the maps of Figures 5 and 6 are defined by single wells. However, high calcium and magnesium values are found in a cluster of wells in southeastern Saskatchewan in all of the four hydrogeological units. Some of these values are greater than the regional exploration limit for calcium and magnesium. In fact, sampled waters from the Dawson Bay Formation in the CanHunter Tableland 6-27-2-9 well show a calcium value of 119,300 mg/l and a magnesium value of 11,420 mg/l, which far exceed those of a commercially produced brine in the Midland Basin. Clustering of anomalously high calcium and magnesium ion concentrations centres around the intersection of latitude 52° N and longitude 107°W for the Souris River/Dawson Bay formations. The Arco Abbey 8-18-21-20W3 well in southwestern Saskatchewan also shows high calcium (93,053 mg/l) and magnesium (11,479 mg/l) values for waters taken from the Souris River Formation.

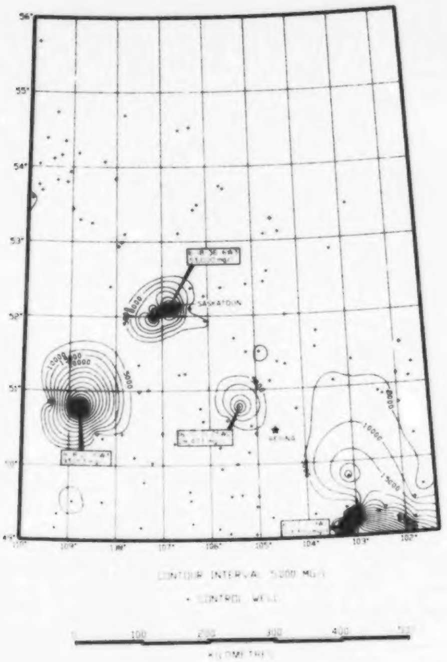
Other sites of anomalous brines are represented by brine inflows into potash mines in the Saskatoon area (south-central Saskatchewan) and in the Rocanville potash mine (southeastern Saskatchewan).

Table 2: Minimum exploration limits for industrial minerals from formation waters (after Ogu and Arnold, 1989).

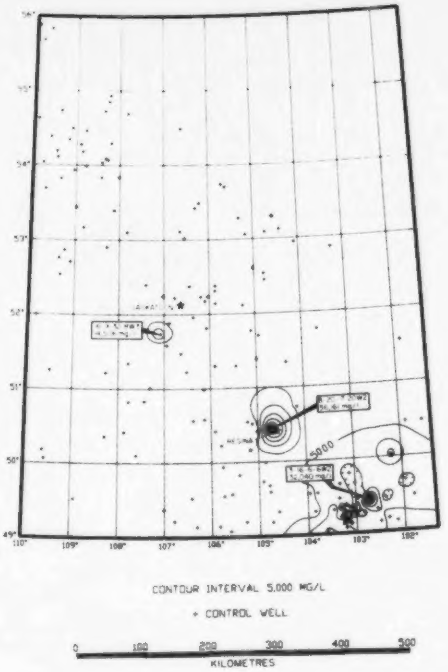
Component	Minimum Regional Exploration Limit (mg/l)	Minimum Detailed Exploration Limit (mg/l)
Calcium	20,000	60,000
Magnesium	3,000	9,000
Bromine	1,000	3,000
Iodine	40	100
Boron	-	-
Lithium	50	-

ECONOMIC POTENTIAL OF SASKATCHEWAN BRINES

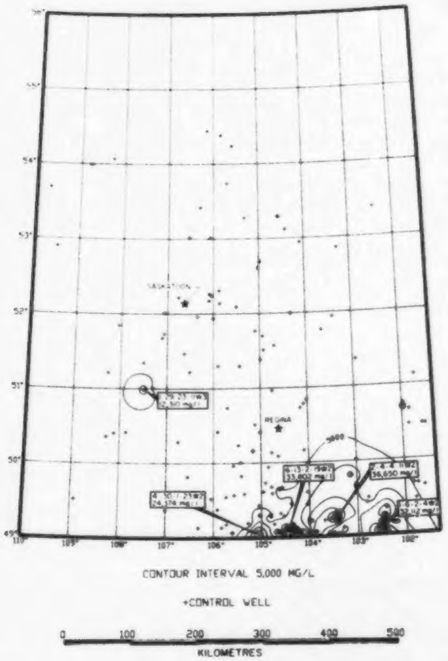
Brines and associated evaporite deposits are the principle source of raw materials for a multi-billion dollar chemical industry in the Michigan Basin of north-central United States. Compounds extracted from brines are the basis for a chemical industry in other parts of the world (*e.g.*, Israel). Brines represent



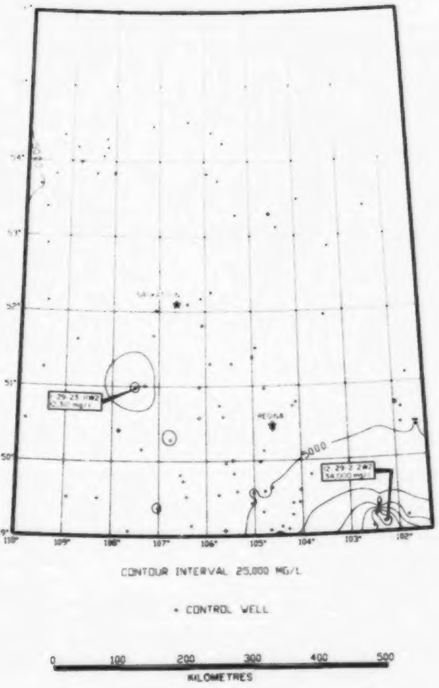
SOURIS RIVER/DAWSON BAY
CALCIUM CONCENTRATIONS



ASHERN/WINNIPEGOSIS/PRAIRIE EVAP
CALCIUM CONCENTRATIONS

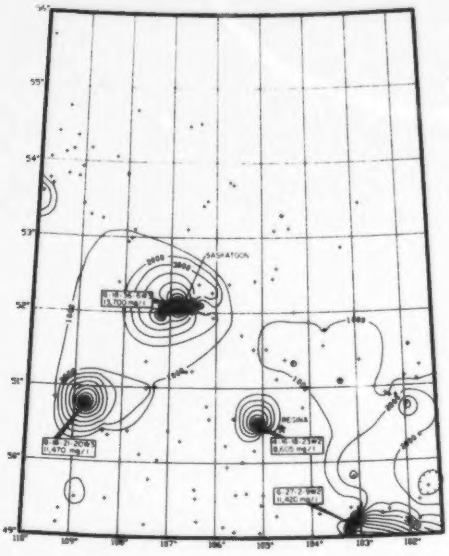


INTERLAKE/STONEWALL/STONY MOUNTAIN
RED RIVER
CALCIUM CONCENTRATIONS



PRECAMBRIAN/EARLIE/DEADWOOD
WINNIPEG CALCIUM
CONCENTRATIONS

Figure 5: Maps showing distribution of calcium in all four hydrogeological units.

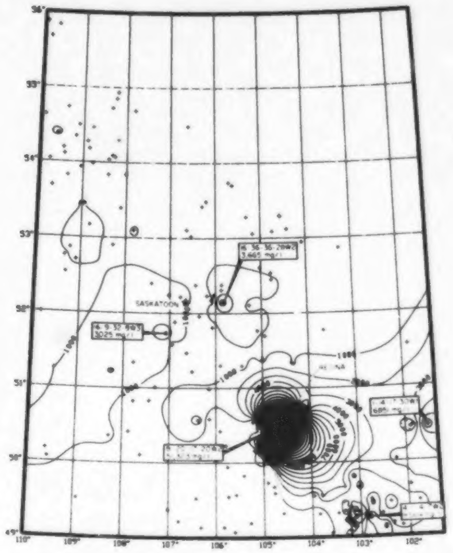


CONTOUR INTERVAL 1,000 MG

+CONTROL WELL

0 100 200 300 400 500
KILOMETRES

SOURIS RIVER/DAWSON BAY
MAGNESUIM CONCENTRATIONS

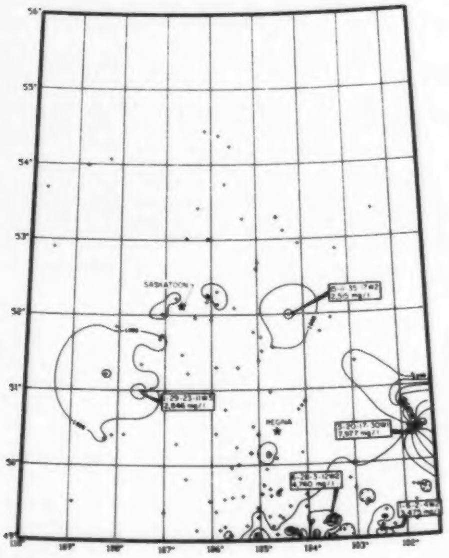


CONTOUR INTERVAL 25,000 MG/L

+CONTROL WELL

0 100 200 300 400 500
KILOMETRES

ASHERN/WINNIPEGOSIS/PRAIRIE EVAP
MAGNESUIM CONCENTRATIONS

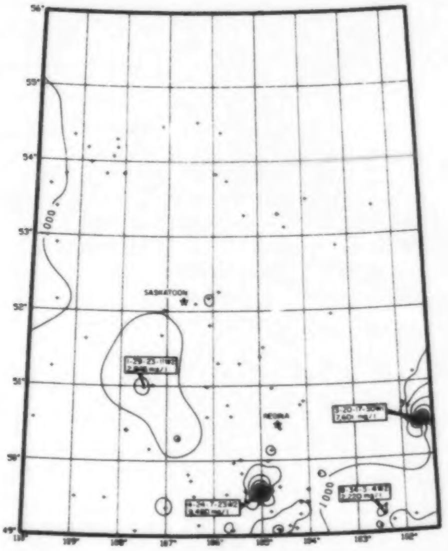


CONTOUR INTERVAL 1,000 MG/L

+CONTROL WELL

0 100 200 300 400 500
KILOMETRES

INTERLAKE/STONEWALL/STONY MOUNTAIN
RED RIVER
MAGNESIUM CONCENTRATIONS



CONTOUR INTERVAL 1,000 MG/L

+ CONTROL WELL

0 100 200 300 400 500
KILOMETRES

PRECAMBRIAN/EARLIE/DEADWOOD
WINNIPEG MAGNESIUM
CONCENTRATIONS

Figure 6: Maps showing distribution of magnesium in all four hydrogeological units.

an easily exploitable commodity requiring minimal extraction (mining) capital with a wide range of potential products (Fig. 7). Ogu and Arnold (1989) have discussed the demand for, and end uses of, marketable chemicals produced from brines.

Chemical analyses of subsurface brine inflows into the potash mines show many of the waters to be anomalously enriched with one or more of the following ions: calcium, magnesium, potassium, bromide, iodide and strontium (Table 1a). Many of the brines far exceed minimum exploration limits (Table 2). Calcium and magnesium anomalies have been identified in all four of the stratigraphic groups defined in this paper. A number of the brine samples taken from the DSTs record waters which closely approximate those in the potash mine inflows with respect to calcium, magnesium, sodium and chloride. However, standard chemical analyses of DST waters rarely include the other elements of economic interest such as bromide, iodide, boron and strontium. It is distinctly possible that these less common components may occur in the brines that resemble the mine-inflow waters in over-all chemistry. The calcium-enriched brines occur throughout the sub-Duperow formations in the southeastern corner of the study area but tend to be richest and most widespread in carbonate reservoirs within the Dawson Bay and Souris River formations, units which immediately overlie the evaporites of the Prairie Formation (Fig. 5).

As part of this study, several drillstem tests that yielded brines rich in calcium and magnesium were analysed to determine aquifer characteristics and possible restrictions on extent of the aquifers (Vigrass, 1991). Some significant parameters for tests, where detailed test data (pressure-time data) are available, are shown in Table 4. The thickness of effective reservoir was obtained from mechanical logs and the average permeability of the reservoir was calculated based on the assumption that the produced fluid had a viscosity of 0.001 Pa.s (1 centipoise). The ideal flow rate is the expected production rate at an arbitrary drawdown of 5 M Pa (510 m of fresh water), when the altered or damaged zone permeability is the same as the reservoir permeability (damage ratio equals unity). For the tests reported in Table 4, the damage ratio is about one to two except for a test at the base of the Deadwood (13-24-02-03W2) where it is 7.3 and the Souris River test in the Arco Abbey 8-18-21-20W3 well where the calculated damage ratio is 9.1. The Arco Abbey well is particularly interesting because of the high average permeability calculated from the drillstem test (770 md.), the large ideal flow rate with damage removed (600 m^3/d), and the ion concentrations shown in Table 1b (calcium = 93, 053 mg/l, magnesium = 11,470 mg/l). In the Arco Abbey well, core analysis indicates an average permeability for the Souris River reservoir of 14 md. Fractures are mentioned in the core analysis report and the test suggests that the porous zone is fractured or there is cavernous porosity within the radius of investigation of the test.

For the tests reported in Table 4, there are no indications of reservoir depletion during the drillstem tests. An indication of the order of magnitude of the radius of investigation can be obtained from the shut-in portion of a drillstem test. For most of the test, the approximate radius of investigation, which is related to the square root of average permeability, ranges from 10 to 60 m. The test on the Souris River reservoir in the Arco Abbey 8-18-21-20W3 well, however, has a calculated radius of investigation of 400 m. This indicates that the 4.5 m zone tested in this well may extend a considerable distance from the tested well and the well may have penetrated an extensive reservoir zone in the Souris River carbonate. Interest in the Arco Abbey 8-18-21-20W3 well was also indicated by Lane (1990) who calculated the brine content in reservoirs (5 percent porosity cut-off) intersected by the well. Lane indicated the possibility of 4.35 x 10^6 tonnes of brine per square mile (2.6 km^2). This quantity of brine would contain 827,000 tonnes of calcium chloride and 144,000 tonnes of magnesium chloride per square mile (based on 19 weight percent $CaCl_2$; 3.3 weight percent $MgCl_2$ in the brine).

Additional parameters on drillstem tests without detailed test data (pressure-time data) are listed in Table 5. The estimated liquid flow rates on test were based on an estimated 90 m of drill-collar fill (0.004 m^3/m capacity) and the remainder of the recovery in drill pipe (0.007 m^3/m capacity). On this basis, three of the tested intervals have small permeability-thickness or are badly damaged. Two Deadwood tests and one Winnipegosis test had estimated flow rates on test in excess of 100 m^3/d and they probably indicate medium to very large permeability in the tested zone. For the tests listed in Table 5, data are insufficient to determine whether or not the reservoirs showed signs of depletion and restriction.

Table 3: Analyses of sea water and commercial brines (from Hitchon and Holter, 1971).

Sample Details	Chemical Constituents (mg/l)								Total Dissolved Solids
	Na	Ca	Mg	Cl	Br	I	SO_4	HCO_3	
Sea water (standard salinity of 3.5%)	10,760	413	1,294	19,353	67	0.06	2,712	142	35,000
Great Salt Lake	91,000	<200	12,400	166,000	-		23,000	-	298,550
Dead Sea (Average)	34,940	15,800	41,960	208,020	5,920	-	540	240	315,040
Lower Devonian (Sylvania Sandstone, Midland, Michigan - The Dow Chemical Company)	22,500	74,800	9,960	208,000	2,910	40	40	-	313,000

Table 4: Parameters for drillstem tests yielding brines rich in calcium and magnesium where detailed test data (pressure-time data) are available.

Well Location	Formation Tested	Effective Thickness (m)	Permeability From Test (md)	Ideal Flow Rate (m^3/d)
13-24-02-03W2	Deadwood	4.9	1.3	30
01-06-02-04W2	Stony Mtn.	6.1	2.6	8.5
06-27-02-09W2	Dawson Bay	8.0	0.4	3.3
07-24-04-05W2	Souris River	5.2	3.9	11
02-04-04-11W2	Stony Mtn.	11.9	5.2	25
08-18-21-20W3	Souris River	4.5	770	600

Table 5: Parameters for drillstem tests yielding brines rich in calcium and magnesium where no detailed test data are available (no pressure-time data).

Well Location	Formation Tested	Effective Thickness (m)	Permeability From Test (md)	Ideal Flow Rate (m^3/d)
12-29-02-02W2	Deadwood	5.0	very large	367
06-13-02-19W2	Stony Mtn.	3.0	small	26
09-34-03-04W2	Deadwood	5.5	medium	101
03-16-06-06W2	Winnipegosis	15.0	large	142
04-16-18-23W2	Dawson Bay	11.6	small	29
04-18-35-08W3	Souris River	10.0	small	11
03-21-35-08W3	Deadwood	insufficient data		

The data in Tables 4 and 5 indicate that some tested zones yielding enriched brines are relatively impermeable. It is distinctly possible that the best reservoir zones were not tested in the well or that the reservoir properties of the tested zone are better elsewhere in the vicinity.

The demand for minerals derived from brines is well demonstrated by examples of annual brine production capacities of companies with producing facilities in other basins. These include: 22,000 to approximately 700,000 tons calcium; 30,000 tons magnesium; 227 to 907 tons iodine minerals; 2 to 124 tons bromine; and 25,000 to 90,000 tons B_2O_3. In Saskatchewan brines, rarer elements like iodine, bromine and boron are found in concentrations similar to or higher than those in commercially exploited brines (Table 1a). With minimum product values ranging from $153/tonne for calcium chloride flake (regular grade) to $950/tonne for bromine, Saskatchewan brines appear to constitute a promising resource, the economic potential of which is enhanced by the possibly high productivity of some of the brine reservoirs.

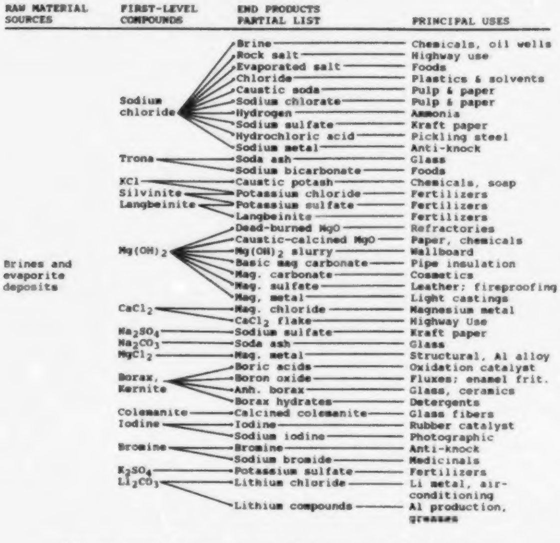


Figure 7. Chemical Tree - Brine and Evaporite Chemicals Industry
(After OGU and ARNOLD, 1989)

Figure 7: Chemical Tree - Brine and evaporite chemicals industry (after Ogu and Arnold, 1989).

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RELATIONSHIPS AMONGST OIL QUALITY, THERMAL MATURITY AND POST-ACCUMULATION ALTERATION IN CANADIAN WILLISTON BASIN (SOUTHEASTERN SASKATCHEWAN AND SOUTHWESTERN MANITOBA)

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ABSTRACT

Oil quality in the northeastern Williston Basin is controlled by source rock characteristics, maturity, and to a lesser extent post accumulation processes, especially biodegradation. Previous studies of Mission Canyon oil quality and compositional variations attributed changes in API gravity, saturate/aromatic hydrocarbon ratio and n-alkane depletion to progressive biodegradation accompanied by meteoric water washing. Source composition and maturity can produce similar effects, but it was previously difficult to characterize oil pool thermal maturity. The trisnorhopane thermal maturity indicator (Ts/Tm) can be used to characterize oil pool maturity and is resistant to biodegradation. Revised oil-source associations quality variations are inferred to be dominated by source rock compositional variations and thermal maturity. Previously proposed gasoline range thermal maturity indicators are of limited value. The few Bakken-sourced oils discovered to date are of uniformly high quality (> 35° API) because of a delayed expulsion threshold (0.9 percent Ro) controlled by source rock composition. These oils are commonly connate water washed, but are not biodegraded. Lodgepole-sourced oils are expelled at the beginning of mainstage hydrocarbon generation (0.7 percent Ro). Their API gravities show a slight sensitivity to thermal maturity that follow progressive changes in saturate/aromatic ratio. Biodegradation, a subsequent and lesser process, overprints a northwestwardly decrease in Madison Group oil pool thermal maturity, but the effects of thermal maturity are inferred to dominate. Some Madison pools are connate water washed without being biodegraded and this probably reflects secondary migration pathway length. Winnipegosis-sourced oils are very sensitive to maturity variations, and oil quality and saturate/aromatic ratio both increase rapidly with maturity. There is no evidence that these oils are biodegraded, although longer migrated oils tend to be more connate water washed. The few discovered Ordovician-sourced pools yield only a limited sample of expected thermal maturities. Ordovician-sourced oil quality is also strongly dependent on thermal maturity. High quality oil pools are expected to be discovered in stratigraphic and diagenetic traps along the eastern flank of the basin, because of the timing and kinetics of Ordovician hydrocarbon generation.

INTRODUCTION

A revision of the oil-source system in southeastern Saskatchewan and southwestern Manitoba allows oil densities, gross compositions and thermal maturities to be examined on a family by family basis. Competing processes, particularly bacterial degradation and thermal maturity can produce similar effects on oil density and composition, and the interpretation of cause can affect how geologists describe the generation, migration and alteration in particular plays.

Darling and Wood (1958) illustrated oil gravity and sulfur content variations in Canadian Williston Basin and showed higher quality oils were geographically associated with the Mississippian depositional centre. API gravity tends to increase toward the southeast along the Mississippian subcrop coincidental with increasing salinity in Mississippian formation waters (Edie, 1958). In the American portion of the eastern Williston Basin important oil density variations occur with stratigraphic position in the Madison Group. Oil density in Mission Canyon stratigraphic traps decreases eastward and downward such that the highest API gravities occur in the lowest stratigraphic zones. In the subcrop reservoirs the reverse is true and oil density increases eastward so that the lowest oil densities (highest oil gravities) are in the higher stratigraphic zones. Improvements in analytical technique and interpretation now allow oil pool thermal maturity estimations, independent of biodegradation and water washing. There is no doubt that biodegradation and water washing affect some oil composition and quality (Bailey *et al.*, 1973b), but its effects have probably been over-emphasized. Thermal maturity and source rock composition are probably more important factors controlling oil quality and composition. An understanding of these would focus exploration for reservoirs that contain the best quality oils.

PREVIOUS WORK

OIL FAMILIES

Williams' (1974) distinguished three stratigraphically restricted oil families by means of carbon isotopic compositions, gasoline range (C_4-C_7) and C_{15+} saturate fraction gas chromatograms (SFGC's). Type I oils, predominantly in Ordovician and Silurian reservoirs, had heavier carbon isotopic compositions, gasoline range n-alkane enrichment and distinctive C_{15+} SFGC's with a strong odd predominance in the range nC_{15}-nC_{20}. Two other families were mutually distinguished using gasoline range nalkane enrichment and SFGC characteristics. Type II oils, the gasoline range of which is relatively depleted in straight chain nalkanes, dominate Upper Devonian, Mississippian and Mesozoic reservoirs. Type III oils, restricted to Pennsylvanian reservoirs, have gasoline range compositions resembling Type I oils, but lacked their distinctive SFGC's. Williams (1974) found considerable overlap of API gravity and sulphur percentage among his three families, but offered no explanation for these variations.

Leenheer and Zumberge (1987) and Zumberge (1983) studied several oil pools, predominantly in the central Williston Basin. They considered many compositional characteristics and concluded that normalized tricyclic terpane distributions (identified by them as tricyclic diterpanes) and carbon isotopic compositions best distinguished oil families. Other compositional characteristics help to define families, but these they considered dependent on maturity *(ibid.)*. Two families have compositions and stratigraphic distributions similar to families defined by Williams (1974). Group 1 oils, including Upper Ordovician Red River pools and the Raymond Middle Devonian Winnipegosis pool, have Type I compositions (Williams, 1974). Other work linked these oils to the Mid-continent (Longman and Palmer, 1987) and global Middle and Upper Ordovician oil-source systems (Hoffman *et al.*, 1987; Foster *et al.*, 1986). Group 2 oils of Leenheer and Zumberge (1987) resemble Type II oils of Williams (1974). Leenheer and Zumberge (1987) did not examine Pennsylvanian reservoirs nor did they find oils compositionally similar to Williams' Type III. Their Group 3 include very mature Duperow Formation pools from the central Williston Basin, and their Group 4 incorporates two Type II immature oils from Upper Devonian reservoirs in the Raymond Field. A single immature oil of peculiar composition from the Deadwood Formation in Newporte Field, North Dakota constitute Group 5.

Brooks *et al.* (1987), in a study of Canadian pools, also found that oils in Ordovician reservoirs, their Family A, resemble Type I (Group 1). They suggested that oils in Devonian, Mississippian and Mesozoic strata, that others would have assigned to Type II or Group 2, be subdivided into at least two families that they inferred to have different sources. Most Mississipppian and Mesozoic pools were assigned to Family C, and were characterized by strong even-odd n-alkane predominance, pristane/phytane (Pr/Ph) < 1, high C_{23}tricyclic/C_{30}pentacyclic terpane ratios (C_{23}/C_{30}) and very low diasterane/regular sterane ratios (DIA/REG). Family B, the remaining Birdbear Formation, Bakken Formation and Madison Group samples, with Pr/Ph > 1, SFGC's without predominance and DIA/REG > 1, were subdivided into Bi, the C_{23}/C_{30} of which resembled that of Family A, and Bii, with C_{23}/C_{30} comparable to or greater than Family C. DIA/REG variations in southeastern Saskatchewan did not match C_{29} sterane maturity variations (Brooks *et al.*, 1987), which lead to the inference that DIA/REG reflected source rock lithology and diagenesis (c.f. Sieskind *et al.*, 1979; Rubinstein *et al.*, 1975). Family C oils were attributed to an euxinic carbonate source, and Family B to a clay-rich clastic source. Family D (Brooks *et al.*, 1988), Winnipegosis pinnacle reef discoveries, extended the stratigraphic range of oils with very low C_{23}/C_{30} and complex SFGC's, like Bi. This is the Family D oil of Brooks *et al.* (1988), and it completely overlaps Group 3 and 4 of Leenheer and Zumberge (1987). Osadetz *et al.* (1990), examined oils in this stratigraphic interval from Saskatchewan and North Dakota and as a result enlarged Family D oils to include oils previously assigned to Family Bi (Brooks *et al.*, 1987).

Osadetz *et al.* (in review) conducted a comprehensive re-evaluation while taking into account new discoveries and oils from Manitoba. They concluded that terpane and SFGC compositional characteristics are adequate for subdividing oil families and ascribed DIA/REG sensitivity to thermal maturity. This resulted in some minor reclassification of oil samples (Osadetz *et al.*, 1990; Brooks *et al.*, 1987; 1988) but retained the four distinctive oil families (Fig. 1). Familial associations are based on C_{23}/C_{30} terpane ratio, prominence of extended homologous hopanes, n-alkane predominance and acyclic isoprenoid compound abundances, particularly pristane and phytane.

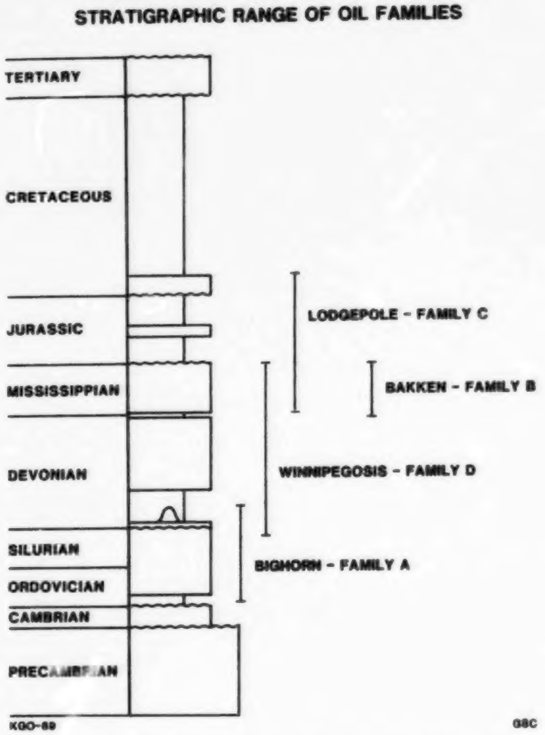

Figure 1: Stratigraphic range of the four oil families identified in this study. Source rock formations for each of the oil families are associated with their oils.

Each compositionally distinctive family has a restricted stratigraphic distribution. Oil pools below the Bakken all have $C_{23}/C_{30} < 0.26$ and a strong 17alpha(H)-tetrakishomohopanes (C_{34} hopanes) prominence (Fig. 2, Table 1). One group among these, occurring most frequently in Ordovician Bighorn reservoirs, but also in a Winnipegosis pool at Minton, has low pristane and phytane abundance, a strong odd-even predominance among n-alkanes between C_{15} and C_{20} and a low relative abundance in higher carbon number n-alkanes (Fig. 3). These oils are compositionally identical to Family A (c.f. Brooks *et al.*, 1987), Group 1 and Type I oils and are referred to as Family A oils. Oils with similar terpane compositions (Fig. 2), but having abundant pristane and phytane (Fig. 3), and complex SFGC's occur primarily in Winnipegosis pinnacle reefs, but also at higher stratigraphic levels such as in the Birdbear Formation and in a number of Ratcliffe beds pools of the Oungre trend. All Family Bi of Brooks *et al.* (1987) and Family D of Osadetz *et al.* (1990) and Brooks *et al.* (1988) are now referred to as Family D oils.

Three oils in the Bakken (#515, #554, #1279), has $C_{23}/C_{30} > 0.40$, lack extended hopane prominence (Fig. 2), and have Pr/Ph > 1.50 and SFGC's lacking predominance (Fig. 3). These are similar to a restricted Family Bii (Brooks *et al.*, 1987) sample set and are now referred to as Family B oils. Many oils that Osadetz *et al.* (in review) studied were from Madison and younger reservoirs and had $C_{23}/C_{30} > 0.30$, a common 17alpha(H)-pentakishomohopane (C_{35} hopane) prominence (Fig. 2), Pr/Ph < 1.1, and an even-odd nalkane predominance (Fig. 3). These include all Family C and some Family Bii oils of Brooks *et al.* (1987), and are combined into a revised Family C.

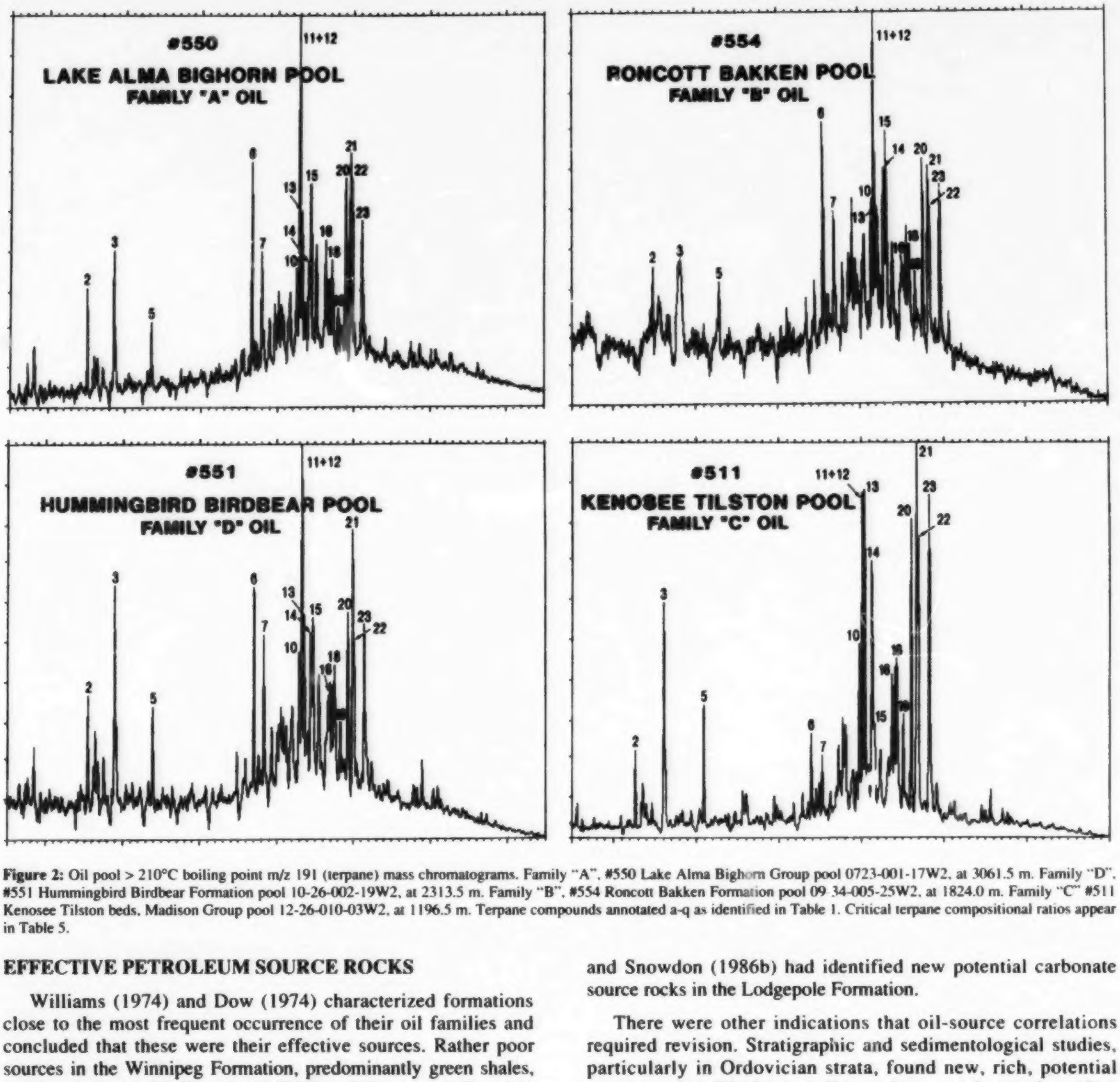

Figure 2: Oil pool > 210°C boiling point m/z 191 (terpane) mass chromatograms. Family "A", #550 Lake Alma Bighorn Group pool 0723-001-17W2, at 3061.5 m. Family "D", #551 Hummingbird Birdbear Formation pool 10-26-002-19W2, at 2313.5 m. Family "B", #554 Roncott Bakken Formation pool 09-34-005-25W2, at 1824.0 m. Family "C" #511 Kenosee Tilston beds, Madison Group pool 12-26-010-03W2, at 1196.5 m. Terpane compounds annotated a-q as identified in Table 1. Critical terpane compositional ratios appear in Table 5.

EFFECTIVE PETROLEUM SOURCE ROCKS

Williams (1974) and Dow (1974) characterized formations close to the most frequent occurrence of their oil families and concluded that these were their effective sources. Rather poor sources in the Winnipeg Formation, predominantly green shales, were associated with Type I; the Bakken shales were believed to have sourced Type II; and the shales of the Tyler Formation were designated the Type III sources (Williams, 1974; Dow, 1974). Subsequent geochemical studies elaborated on source potential and effectiveness within the framework of these correlations and often emphasized Bakken sources (Dembicki and Pirkle, 1985; Price *et al.*, 1984; Lenheer, 1984; Webster, 1984; Schmoker and Hester, 1983; Meissner, 1978). In fact, one study (Thode, 1981) suggested that Type II oils were exclusively sourced by the Lower member of the Bakken Formation.

Leenheer (1984) showed that immature Bakken extracts had DIA/REG < 1, like the least mature Group 2 oil (Leenheer and Zumberge, 1987) and most Family C oils (Brooks *et al.*, 1987). This observation and their conclusion (Leenheer and Zumberge, 1987) that DIA/REG varied continuously with thermal maturity strengthened the association between Bakken sources and Group 2 oils. This differed from Brooks *et al.* (1987), who used DIA/REG to propose a carbonate source for Family C oils, because Osadetz and Snowdon (1986b) had identified new potential carbonate source rocks in the Lodgepole Formation.

There were other indications that oil-source correlations required revision. Stratigraphic and sedimentological studies, particularly in Ordovician strata, found new, rich, potential sources using lithology, sedimentology, total organic carbon (TOC) abundance and anhydrous pyrolysis (Rock-Eval) (Osadetz et al., 1989; Osadetz and Snowdon, 1986b; Kohm and Louden, 1982, 1978; Kendall, 1976; Thomas, 1968). Additional Winnipegosis discoveries and recognition of regionally important sources in the Winnipegosis Formation (Osadetz and Snowdon, in review; Osadetz *et al.*, 1990; Stoakes *et al.*, 1987; Rosenthal, 1987; Wardlaw and Reinson, 1971) could not be accommodated by existing oil-source models.

Osadetz and Snowdon (in review) found four regionally significant Paleozoic source rock formations. These include, 1) Type I kukersites in Upper Ordovican Bighorn Group, and Type II sources in, 2) Middle Devonian Upper Winnipegosis starved basin facies, 3) Upper Devonian-Mississippian Bakken shale members, and 4) bituminous carbonate rock bodies within the Mississippian Lodgepole Formation. Lesser, or less well described sources, were found at undo-beds maximum flooding surfaces in the Winnipeg and Winnipegosis formations, and in

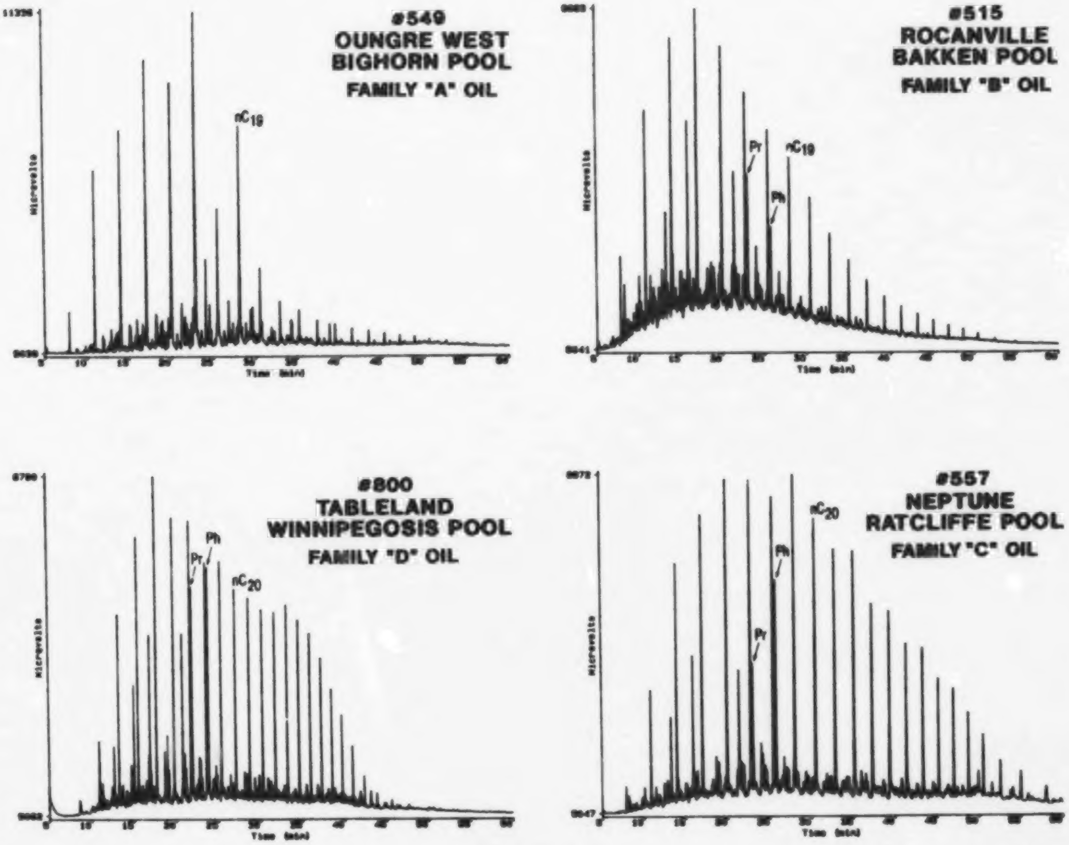

Figure 3: Oil pool > 210°C boiling point Saturate Fraction Gas Chromatograms. Family "A", #549 Oungre West Bighorn Group pool 04-22-003-15W2, at 2932.5 m. Family "D", #800 Tablelands Winnipegosis reef 15-02- 003-10W2, at 2562.0 m. Family "B", #515 Rocanville Bakken Formation pool 02-11-016-31W1, at 654.3 m. Family "C" #557 Neptune Ratcliffe beds, Madison Group pool 04-06- 004-16W2, at 1752.0 m. Pristane/phytane ratios as listed in Table 5.

TABLE 1: Saturate fraction biological marker compounds identified as peaks on m/z 191 terpane traces

a — C23 tricyclic terpane
b — C24 tetracyclic terpane
c — Ts [18alpha(H)-trisnorhopane]
d — Tm [17alpha(H)-trisnorhopane]
e — 17alpha(H)-norhopane
f — 17alpha(H)-hopane, C30 pentacyclic terpane
g — moretane
h — (20S) 17alpha(H)-homohopane, C31 pentacyclic terpane
i — (20R) 17alpha(H)-homohopane, C31 pentacyclic terpane
j — (20S) 17alpha(H)-bishomohopane, C32 pentacyclic terpane
k — (20R) 17alpha(H)-bishomohopane, C32 pentacyclic terpane
l — (20S) 17alpha(H)-trishomohopane, C33 pentacyclic terpane
m — (20R) 17alpha(H)-trishomohopane, C33 pentacyclic terpane
n — (20S) 17alpha(H)-tetrakishomohopane, C34 pentacyclic terpane
o — (20R) 17alpha(H)-tetrakishomohopane, C34 pentacyclic terpane
p — (20S) 17alpha(H)-pentakishomohopane, C35 pentacyclic terpane
q — (20R) 17alpha(H)-pentakishomohopane, C35 pentacyclic terpane

other formations with local or undefined potential (Campbell and Forbes, 1990; 1989).

REVISED OIL-SOURCE CORRELATION

Stratigraphically suggested oil-source associations were generally corroborated by compositional similarities among extracts and oils (Osadetz *et al.*, in review). Bighorn kukersite solvent extracts, like Family A oils, have distinctive SFGC's, very low C_{23}/C_{30}, and a C_{34} hopane prominence at higher maturities. Rich sources from the Winnipeg Formation in Canada do not have the characteristic kukersite SFGC, although American extracts do (Williams, 1974, his Fig. 13), so that several formations of the Bighorn Group and the Winnipeg Formation are possible sources for Family A oils (Fig. 1). Both Winnipegosis and Lodgepole extracts have low C_{23}/C_{30} terpane and complex SFGC's like Family D oils, but Lodgepole extracts have a C_{35} hopane prominence and a consistent even n-alkane predominance, like Family C oils. Some Winnipegosis sources and all Family D oils have C_{34} hopane prominences such that the Winnipegosis is inferred to be the effective source for all Family D oils (Fig. 1). Platform Winnipegosis source rocks are not well known (Osadetz and Snowdon, in review) nor how different Winnipegosis source rock facies contribute to Family D oils. Bakken shales source Family B oils (Fig. 1). Bakken extracts, like Family B oils, have consistently higher C_{23}/C_{30} and Pr/Ph ratios, lack extended hopane prominence and show no n-alkane predominance. Bakken extract compositional variation is much greater than that characterizing Bakken oils, thereby indicating that only a restricted portion of the unit is the effective oil source. Compositional similarity amongst Family C pools implies a single source and solvent extracts from Lodgepole sources west of Range 20W2 (well west of the oil fields), have low Pr/Ph, strong even-odd n-alkane predominance, C_{35} hopane prominence and low DIA/REG like Family C oils at appropriate maturities. These or similar Lodgepole rocks are the sources of Family C oils (Fig. 1), however, their C_{23}/C_{30} ratios do not match those of the oil pools, but this may be due to primary migration effects (Seifert *et al.*, 1980).

Additional oil families are known and they probably have distinctive sources. Type III — Tyler Formation oils (Williams, 1974) are one example. Not all the compositionally distinctive oils described by Leenheer and Zumberge (1987) and Zumberge (1983) have identified sources, while Grantham and Wakefield (1988) have suggested that effective oil sources occur in the Mission Canyon, Red River and Duperow formations. Whether or not any of these oil families or sources are equivalent to those proposed by Osadetz *et al.* (in review) has not been determined.

WATER WASHING, BIODEGRADATION AND THERMAL MATURITY

Darling and Wood (1958) speculated that both sulphur and gravity varied due to source and "environment" and that these might have been related to "minor basin tectonics" (ibid.). Edie (1958) did not provide a mechanism for the association of water salinities and crude oil gravity variations. By the early 1970's a number of studies conducted or sponsored by Imperial oil described the importance of post-accumulation processes on oil quality (Evans *et al.*, 1971; Bailey *et al.*, 1973a; 1973b). Jobson *et al.* (1972) had experimented with the effects of bacterial cultures on a North Cantal field oil and found that its specific gravity could be altered from 0.827 (40° API) to 1.046 (5° API) in 21 days. Bailey *et al.* (1973b) made comprehensive studies of compositional and isotopic changes of MC_3 zone (Frobisher Beds) crude oils and formation waters. They assumed that all of their Frobisher beds oils belong to the same family (Type II) and this remains unchanged by subsequent revision, although their samples would now be assigned a Family C — Lodgepole affiliation. They observed changes in gross and molecular compositions, but found the most pervasive alteration to be in the gasoline range (C_4-C_7). The selective and progressive removal of the most soluble compounds, benzene and toluene, was inferred to be indicative of water washing. Relative n-alkanes depletion with respect to branched and cyclic compounds in both the gasoline range and C_{15+} compositional fraction was inferred to indicate biodegradation. Progressive n-hexane and n-heptane removal was particularly noticable while nC_{18}/phytane ratios changed progressively in the heavier fraction. Other progressive changes were noticeable in lighter napthenes, saturate/aromatic ratio, gas/oil ratio, sulphur content and API gravity (ibid.).

Carbon isotopic variations for various compositional fractions were slight (Bailey *et al.*, 1973b), comparable to unaltered Group 2 oils and Bakken extracts (Leenheer and Zumberge, 1987), but did increase slightly northwestward towards increasingly biodegraded oils. Thode (1981) examined sulphur isotopes in the same pools and observed that they generally paralleled API gravity, chemical composition and formation water salinity variations. Sulphur isotopic compositions changed more significantly than expected in the more southeasterly oil pools (inferred to be less biodegraded). This led to the deductions that sulphur is being derived from reservoir sulphate by anaerobic microbial action and that there is an isotopic exchange between sulphur compounds in the oil and hydrogen sulfide in the reservoir. Several other compositional characteristics, notably nC_{17}/pristane, did not exhibit progressive variations commonly attributed to biodegradation.

Similar variations are also commonly attributed to low thermal maturity (Hunt, 1979; Montgomery, 1974). Bailey *et al.* (1973b) had no thermal maturity indicator to show that compositional variations were independent of maturity. Subsequent studies inferred maturity from isotopic (Thode, 1981) and gasoline range (Thompson, 1983) compositions. Both studies used oil families defined by Williams (1974). Thompson inferred long lateral migration pathways and uniformly high maturities, VR = 1.1 percent, that he concluded were an indication of a delayed expulsion threshold. Thode (1981) concluded that certain Type I and II (c.f. Williams, 1974) oils had been thermally altered, although he did not relate these variations to standard maturities.

Snowdon and Osadetz (1988a, 1988b) studied gasoline range (iC_5-nC_8) compositions to determine if familial affiliations proposed by Brooks *et al.* (1987) were reflected in that compositional fraction. Their observations were comparable to Williams's (1974), but they concluded that most gasoline range variations were not attributable to source or thermal maturity. Considerable compositional variation was attributed to both water washing and biodegradation (Fig. 4). Like Bailey *et al.* (1973b) they found that both water washing and biodegradation were highest at the northwestern limit of the Mississippian oilfield trend, but they also found that eastern portions of the producing area were significantly water washed without being biodegraded. Osadetz *et al.* (in review) proposed a new relationship between the trisnorhopane ratio (Ts/Tm) and thermal maturity that allows analysis of how composition varies with maturity.

EXPERIMENTAL METHODS

Approximately 60 oil pools in Saskatchewan and Manitoba were examined. Analysis of the gasoline range compositional

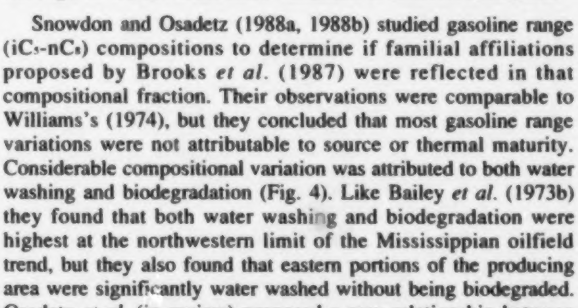

Figure 4: Factor analysis Factor 2 and Factor 3 scores for gasoline range compounds from (Snowdon and Osadetz, 1988b). In Factor 2 volatiles, benzene and toluene had high negative loadings that were interpreted to indicate water washing. In Factor 3 benzene, toluene and branched compounds had high positive loadings while normal alkanes had high negative loadings. Factor 3 was intepreted to reflect n-alkane enrichment or depletion that was either characteristic of Family A and D affinities or suggestive of biodegradation.

Table 2: Gasoline Range Compounds

1. iC5 - isopentane
2. nC5 - n-pentane
3. 22DMC4 - 2,2-dimethylbutane
4. CYC5 - cyclopentane
5. 23DMC4 - 2,3-dimethylbutane
6. 2MC5 - 2-methylpentane
7. 3MC5 - 3-methylpentane
8. nC6 - n-hexane
9. MCYC5 - methylcyclopentane
10. PHI - benzene
11. CYC6 - cyclohexane
12. 2MC6 - 2-methylhexane
13. 11DMCYC5 - 1,1-dimethylcyclopentane
14. 23DMC5 - 2,3 dimethylpentane
15. 3MC6 - 3-methylhexane
16. 1C3DMCYC5 - 1cis,3-dimethylcyclopentane
17. 1T3DMCYC5 - 1trans,3-dimethylcyclopentane
18. 1T2DMCYC5 - 1trans,2-dimethylcyclopentane
19. nC7 - n-heptane
20. 1C2DMCYC5 - 1cis,2-dimethylcyclopentane
21. MCYC6 - methylcylohexane
22. 25DMC6 - 2,5-dimethylhexane
23. 24DMC6 - 2,4-dimethylhexane
24. 223TMC5 - 2,2,3-trimethylpentane
25. PHIC - toluene
26. 3MC7 - 3-methylheptane
27. 1C4DMCYC6 - 1cis,4-dimethylcyclohexane
28. nC8 - n-octane

fraction (iC_5-nC_8) followed techniques used by Snowdon and Osadetz (1988a,b). Briefly, a whole crude oil aliquot was mixed with an inert solid (water saturated alumina or Ottawa sand) and placed into a sample chamber through which a fixed volume of helium was passed at room temperature. The effluent was then passed onto a piece of fused-silica chromatography column held at the temperature of liquid nitrogen. When the liquid nitrogen was removed the gasoline range hydrocarbons were allowed to proceed onto a 60 m DB-1 fused silica chromatography column for analysis using a flame ionization detector and Nelson Analytical integration system.

Gross compositions > 210°C were determined using various fractions obtained from packed column elutions following techniques similar to those discussed by Snowdon (1978). The > 210°C boiling point saturate fractions were examined by capillary column gas chromatography (GC) using a VARIAN 3700 gas chromatograph equipped with a 25 m fused-silica column coated with OV-1, temperature programmed from 50-280°C at 4°C/min. and using He as the carrier gas.

C_{15+} saturate fraction capillary column gas chromatography-mass spectrometry (GC-MS) was performed using either a KRATOS MS-80RF mass spectrometer controlled by a DS-55 data system or a VG 70SQ hybrid gas chromatograph-mass spectrometer-mass spectrometer (GC-MS-MS) in GC-MS mode. GC-MS conditions were typically: J and W 25 m DB5 column

Table 3: Oil Sample Gasoline Range Compositional Characteristics and API Gravities

#	PI1	HV	Q	R	S	T	U	V	W	B	API
FAMILY A											
549	0.96	59.7	4.26	3.49	1.58	5.33	3.88	0.17	0.08	0.05	28.4
550	1.14	59.3	4.75	3.21	4.73	67.04	14.08	0.18	0.06	0.00	33.3
920	0.68	34.0	1.81	1.32	3.13	36.08	6.13	0.37	0.17	0.02	24.0
1165	0.78	58.4	6.40	3.12	2.00	15.95	2.26	0.39	0.10	0.02	33.0
FAMILY B											
515	0.52	22.3	1.95	1.47	54.00	0.64	13.00	0.80	0.47	1.06	36.5
554	0.45	24.2	1.21	1.49	9.44	10.99	4.64	0.51	0.39	0.06	40.5
1279	0.71	20.1	1.39	1.30	446.20	178.20	210.30	0.48	0.43	0.00	40.0
FAMILY C											
494	0.55	22.5	1.18	1.25	2.36	1.38	2.24	0.57	0.50	0.58	27.0
495	1.19	25.1	1.86	0.95	148.40	28.90	436.10	0.55	0.33	0.04	36.8
497	0.84	26.7	1.22	1.23	0.39	12.26	0.96	0.57	0.37	0.70	30.6
499	1.07	27.8	1.89	0.98	0.66	3.90	2.35	0.41	0.25	0.26	36.4
500	1.85	27.7	2.02	1.29	0.46	5.25	1.04	0.66	0.41	0.15	30.2
511	0.76	26.4	1.57	1.21	1.57	5.68	3.34	0.54	0.35	0.15	35.4
512	0.90	0.8	0.11	0.30	16.00	42.14	83.00	5.00	1.33	0.88	17.8
514	0.49	22.9	0.99	1.34	0.57	1.18	0.63	0.53	0.43	0.63	32.5
516	1.12	27.5	1.79	0.97	1.72	0.91	7.17	0.52	0.27	1.12	37.6
520	1.39	28.7	1.99	1.15	0.34	0.69	1.14	0.63	0.31	1.27	36.2
523	0.22	1.5	0.23	0.06	250.70	16602.00	577.91	1.35	1.37	0.04	24.7
533	19.10	18.8	0.85	0.83	26.80	588.90	40.40	0.82	0.68	0.00	32.6
539	0.80	22.9	1.48	1.25	0.63	1.41	1.06	0.63	0.41	0.57	37.8
540	0.63	21.7	1.09	1.20	0.41	0.76	0.59	0.59	0.44	1.10	28.4
547	1.42	25.4	2.01	1.14	0.43	1.10	1.02	0.46	0.36	0.80	36.8
548	1.61	23.5	2.14	0.67	1.26	2.22	3.11	0.59	0.35	0.67	33.4
553	0.63	25.4	1.34	1.09	2.13	1.61	1.29	0.39	0.38	0.37	
555	0.60	27.3	1.11	1.40	0.26	1.01	0.51	0.41	0.33	0.71	33.4
556	0.65	27.0	1.29	1.35	0.25	0.64	0.40	4.69	0.37	1.15	31.1
557	0.95	27.6	1.19	1.18	0.29	0.66	0.88	0.63	0.39	1.29	24.0
559	0.66	25.6	1.07	1.20	0.30	0.54	0.60	0.54	0.43	1.53	30.4
560	0.63	20.7	0.87	0.84	0.50	1.06	1.40	0.66	0.52	1.13	
565	1.11	26.0	1.97	1.06	3.01	3.05	7.88	0.51	0.33	0.31	36.6
585	0.71	24.1	1.36	1.31	0.86	3.16	1.04	0.63	0.45	0.24	28.2
595	0.96	21.4	1.38	0.88	18.88	12.63	37.44	0.79	0.51	0.09	31.0
596	0.34	7.2	0.48	0.30	37.00	8.59	114.60	0.50	0.63	0.39	24.2
711	0.91	24.8	1.62	1.03	1.53	4.17	4.93	0.47	0.30	0.23	35.6
714	0.81	25.6	1.55	1.07	0.86	3.45	2.35	0.53	0.32	0.27	33.2
717	0.74	17.6	1.47	0.81	2.82	5.02	6.87	0.52	0.37	0.25	36.6
721	1.12	29.3	2.13	1.23	5.00	51.71	11.92	1.22	0.31	0.02	33.2
722	0.98	14.3	1.14	0.51	29.69	269.86	83.15	0.69	0.53	0.00	34.6
725	1.03	16.5	1.28	0.57	22.11	393.15	57.89	0.65	0.52	0.00	33.0
FAMILY D											
466	0.60	25.6	1.46	1.65	8.45	506.00	4.39	0.58	0.35	0.00	33.7
546	0.60	21.1	1.18	1.12	1.45	2.66	1.94	0.59	0.45	0.34	31.0
551	0.41	22.3	1.01	1.25	1.38	1.50	1.31	0.44	0.37	0.53	39.0
552	0.64	31.1	1.87	1.46	0.36	0.86	0.27	0.33	0.27	0.80	36.0
558	0.78	25.7	0.99	0.79	0.80	0.37	0.49	0.52	0.29	3.47	29.0
568	0.81	22.0	61.57	0.85	0.29	2.51	0.81	0.95	0.40	0.47	30.0
577	0.70	21.7	0.94	0.68	0.26	1.82	1.19	0.59	0.44	0.81	32.0
755	1.36	28.8	39.47	1.13	0.43	2.51	0.70	0.56	0.29	0.35	35.6
756	1.51	36.9	2.44	1.42	0.58	4.21	1.17	0.51	0.22	0.17	39.3
800	1.26	27.6	1.84	1.01	0.50	3.12	0.78	0.23	0.30	0.32	34.0
802	0.92	24.2	1.39	0.99	0.92	6.01	1.20	0.59	0.39	0.17	33.0
841	0.84	18.4	0.88	0.79	0.79	2.93	0.73	1.00	0.55	0.43	24.9
842	0.80	18.8	0.91	0.74	0.96	2.50	0.97	0.86	0.55	0.54	24.9
924	1.26	25.0	1.62	1.04	0.94	2.93	1.28	0.50	0.33	0.33	34.0
1275	1.99	40.8	3.20	1.85	0.75	4.66	1.02	0.47	0.20	0.12	40.5
1276	1.97	27.6	2.93	1.14	2.41	20.03	2.48	0.49	0.27	0.04	40.5

LEGEND TO COLUMNS IN TABLE 3:

# - sample number in Table 4

PI1 - PARAFFIN INDEX 1 = peaks (12+15)/(16+17+18)

HV - HEPTANE VALUE = peaks 19/(sum of peaks 11 to 21)

Q = nC_6/MCYC5

R = nC_7/MCYC6

S = $3MC_5$/BENZENE

T = MCYC6/TOLUENE

U = CYC6/BENZENE

V = iC_5/nC_5

W = $3MC_5/nC_6$

B = TOLUENE/nC_7

API = API gravity in degrees

Table 4: Sample location descriptions

GSC#	POOL NAME	DEPTH (m)	LOCATION	ZONE
0466	NO POOL NAME	1080.3	05-33-010-32W1	Birdbear
0494	INNES WEST	1335.4	13-31-007-11W2	Frobisher
0495	SOUTH PARKMAN	1090.8	16-09-008-33W1	Til/Souris V.
0497	VIEWFILED	1349.8	13-33-007-08W2	Frobisher
0499	STEELMAN	1394.1	13-27-004-05W2	Midale
0500	BENSON	1335.3	05-25-006-08W2	Midale
0511	KENOSEE	1196.5	12-26-010-03W2	Tilston
0512	RED JACKET	664.5	10-04-014-32W1	Jurassic
0514	RALPH	1308.5	01-32-007-13W2	Midale
0515	ROCANVILLE	654.3	02-11-016-31W1	Bakken
0516	NOTTINGHAM	1134.6	04-14-005-33W1	Alida
0520	WORKMAN	1212.5	03-03-002-32W1	Frobisher
0523	WAPELLA	670.2	13-34-014-01W2	Wapella
0533	STAR VALLEY	1161.5	13-11-009-06W2	Frobisher/Alida
0539	ALIDA WEST	1235.1	06-06-006-33W1	Alida
0540	OUNGRE	1823.3	03-29-002-14W2	Ratcliffe
0546	FREDA LAKE	1703.3	16-30-004-18W2	Ratcliffe
0547	ELMORE	1180.3	09-04-001-31W1	Midale
0548	GAINSBOROUGH	1064.0	07-28-002-30W1	Frobisher/Alida
0549	OUNGRE WEST	2932.5	04-22-003-15W2	Red River
0550	LAKE ALMA	3061.5	07-23-001-17W2	Red River
0551	HUMMINGBIRD	2313.5	10-26-002-19W2	Birdbear
0552	KISBEY	1620.5	07-27-007-06W2	Birdbear
0553	HUMMINGBIRD	2226.0	05-26-002-19W2	Bakken
0554	RONCOTT	1824.0	09-34-005-25W2	Bakken
0555	CLAIRLAW	1231.5	05-29-007-05W2	Frobisher/Alida
0556	HUMMINGBIRD	1890.0	11-26-002-19W2	Ratcliffe
0557	NEPTUNE	1752.0	04-06-004-16W2	Ratcliffe
0558	HOFFER	1940.0	05-30-001-15W2	Ratcliffe
0559	LAKE ALMA	1984.0	04-29-001-17W2	Ratcliffe
0560	FREEMANTLE	1189.7	11-16-008-03W2	Watrous
0565	STORTHOAKS	1041.5	A09-17-005-31W1	Alida Tilston
0568	HOFFER	1944.5	09-27-001-15W2	Ratcliffe
0577	FLAT LAKE	1975.0	06-05-001-16W2	Ratcliffe
0585	HANDSWORTH	1178.4	A09-36-010-08W2	Alida
0595	PARKMAN	1082.5	08-11-010-01W2	Souris Valley
0596	MOOSE VALLEY	1183.5	12-14-012-06W2	Tilston
0711	WASKADA	921.0	16-13-001-26W1	Miss. Canyon 1B
0714	WASKADA	928.0	05-03-002-26W1	Miss. Canyon 3B
0717	WASKADA	925.5	04-25-001-26W1	Lower Amaranth
0721	DALY	725.7	10-13-010-28W1	Lodgepole A
0722	VIRDEN	620.5	09-26-010-26W1	Lodgepole B
0725	VIRDEN	607.5	04-27-011-26W1	Lodgepole A
0755	TABLELAND	2601.0	04-36-002-10W2	Winnipegosis
0756	TABLELAND	2581.0	08-22-002-09W2	Winnipegosis
0800	TABLELAND	2562.0	15-02-003-10W2	Winnipegosis
0801	MACOUN	2468.0	16-11-004-10W2	Winnipegosis
0802	BIENFAIT	2436.0	12-31-003-06W2	Winnipegosis
0841	OXBOW	2327.0	10-24-002-03W2	Winnipegosis
0842	OXBOW	2310.0	10-24-002-03W2	Winnipegosis
0902	WEIR HILL	2470.0	09-29-006-06W2	Ordovician
0924	TABLELAND	2607.0	11-30-002-09W2	Winnipegosis
1165	MINTON	2608.9	11-02-003-21W2	Winnipegosis
1275	MACOUN	2380.0	16-30-004-08W2	Winnipegosis
1276	MACOUN	2395.0	06-30-004-08W2	Winnipegosis
1279	DALY	879.5	12-21-018-29W1	Bakken A
6979	MINTON	2863.8	11-02-003-21W2	Bighorn
7595	BEAUBIER	3056.9	03-20-002-16W2	Bighorn

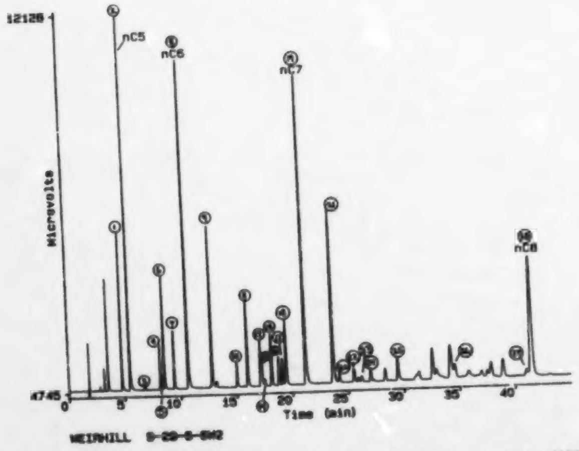

Figure 5: Gasoline range gas chromatogram for the Weirhill pool sample (#920) from 09-29-006-06W2. Gasoline range compounds identified by numbers are identified in Table 2.

temperature programmed from 50-320°C at 4°C/min. and coupled directly to the mass spectrometer source, operated with a 300 microA filament emission current and 70 eV ionization potential. Data were collected by either full mass spectra scanning or by multiple ion detection (MID) monitoring ions at m/z 177.1638, 191.1794, 217.1950, 218.2028, 231.2106, and 259.2262.

RESULTS

Gasoline range compositional characteristics of oil pools (Table 2; Fig. 5) are reported in Table 3. Oil pool (Table 4) > 210°C boiling point gross and saturate molecular compositions characteristics are reported in Table 5. Representative SFGC's (Fig. 3), terpane (Fig. 2) and sterane (Fig. 6; Table 6) mass chromatograms for each oil family are also displayed. Different compositional fractions analyzed by different techniques are commonly given distinctive significance. Compositional variations commonly arise due to differences in source rock kerogen composition attributable to ecological differences in communities of contributing organisms (Hoffmann *et al.*, 1987), depositional environment (e.g. Peters and Moldowan, 1991), mineral matrix (Shimoyama and Johns, 1972) and diagenetic reactions (Sieskind *et al.*, 1979). Other compositional variations arise from source rock thermal maturity (Curiale *et al.*, 1989) and alteration in the reservoir (Bailey *et al.*, 1974).

API GRAVITY (DENSITY) VARIATIONS

API gravities are compiled from provincial and industrial sources. Gravity measurements often vary according to sampling conditions. An example is taken from the Oungre West pool (#549, 04-22-003-15W2) (F. Haidl, pers. comm.). A small oil sample from drill stem test number one (1985/07/11) was reported at 27.9° API. A sample recovered after swabbing (1985/09/11) was 24.7° and a production test sample of the interval 2945.5-2949.5 m (1985/09/16) was reported at 24.49°. Subsequently, the well was subsequently completed over the interval 2938-2939.6 m and samples taken after production of 8.7 m^3 (1985/09/24) and 17.3 m^3 (1985/09/25) were both reported as 28.44°. A later sample taken from storage tanks after an acid squeeze (1985/10/28) was reported at 25.6°.

Sample API gravities (Table 5) vary from approximately 17° to 44°, which is a slightly larger range than that studied by Bailey *et al.* (1973b; 15.2° to 37.6°). Distinctive oil families have different, but often overlapping ranges. Family A oils, from Ordovician sources, vary from 24° to 34° API. Family B oils are limited to between 36.5° and 40.5° API, although they are widely distributed both geographically and in depth. Family C oils range between approximately 17° and 37° API, although most are > 26° API. There is a tendency for Family C API gravities to increase southeasterly along the subcrop trend. A number of higher gravity oils, up to approximately 41° API occur in the Madison Group (Saskatchewan Energy and Mines, 1988; Edie, 1958) and these too have Family C affinities (Snowdon and Osadetz, 1988a), although their critical molecular compositions have not been studied. Family D oils have a wide stratigraphic distribution and a range of gravities, from 25° to 44° API (Martindale *et al.*, this volume).

GENERAL COMPOSITIONAL CHARACTERISTICS

The gross compositions of boiling point fraction > 210°C (Table 5) are characterized by percentage of hydrocarbons (HC) compared to resins and asphaltenes, and these range between 36.5 and 95.6 percent, with both the highest and lowest values belonging to Family D. Family A oils do not reach 90 percent HC and Family B oils are consistently above 90 percent HC. Boiling

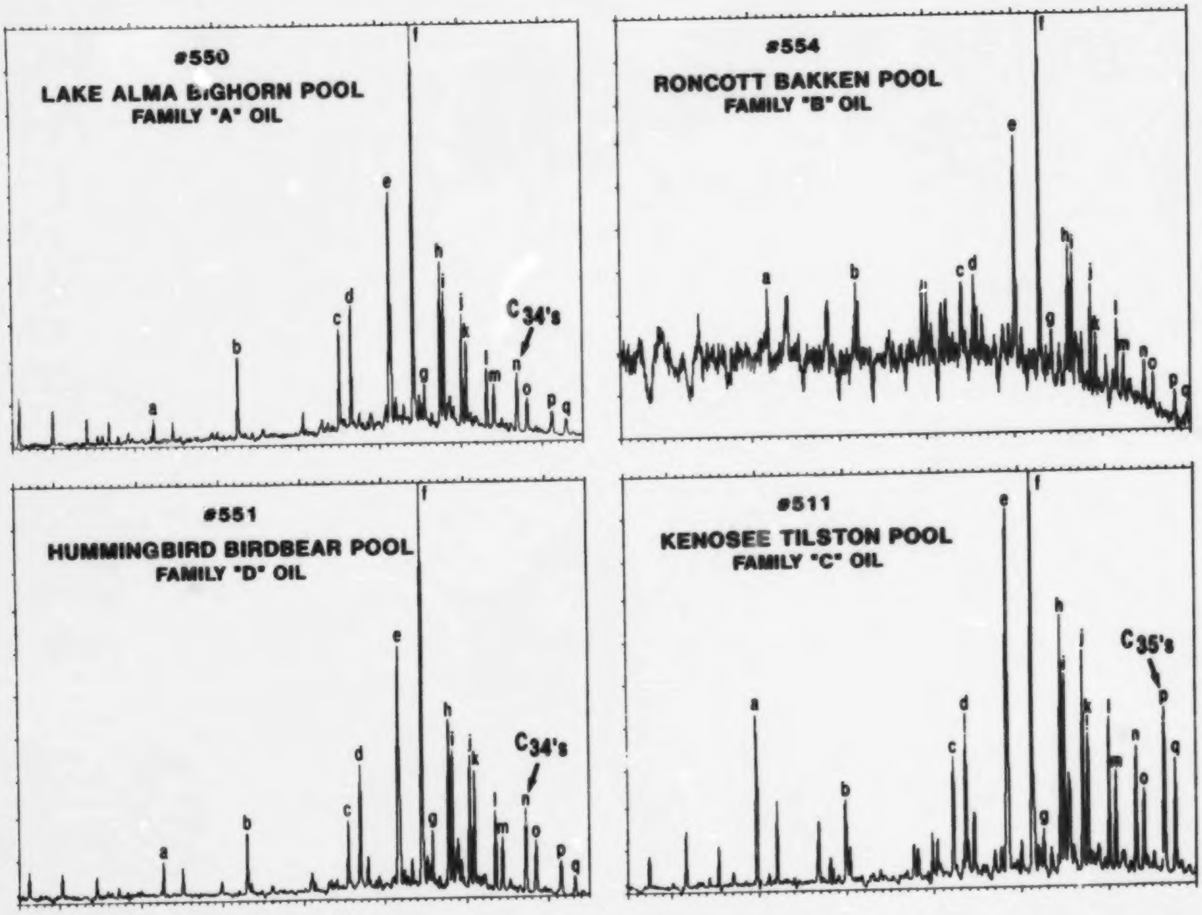

Figure 6: Oil pool > 210°C boiling point m/z 217 (sterane) mass fragmentograms. Family "A", #550 Lake Alma Bighorn Group pool 0723-001-17W2, at 3061.5 m. Family "D", #551 Hummingbird Birdbear Formation pool 10-26-002-19W2, at 2313.5 m. Family "B", #554 Roncott Bakken Formation pool 09-34-005-25W2, at 1824.0 m. Family "C" #511 Kenosee Tilston beds, Madison Group pool 12-26- 010-03W2, at 1196.5 m. Sterane compounds annotated 2-23 as identified in Table 6. Critical sterane compositional ratios are listed in Table 5.

point fraction > 210°C saturate/aromatic hydrocarbon ratios vary considerably from approximately 0.6 to approximately 3.4. The widest range is again characteristic of Family D and the narrowest range is shown by Family B (1.16-1.48).

All the oils studied have prominent 210°C SFGC n-alkane peaks (Fig. 3), C17/pristane > 1 and C18/phytane > 1 (Snowdon and Osadetz, 1988a) and > 210°C SFGC n-alkane often have predominances that are important for familial affiliation (Fig. 3, Osadetz *et al.*, in review). High Prairie oil (Bailey *et al.*, 1973b) is the only reported sample from this part of the basin where n-alkane peaks have been severely depleted with respect to pristane and phytane (nC_{17}/pristane and nC_{18}/phytane << 1). The susceptibility of compound classes to biodegradation is generally accepted to be: n-alkanes > acyclic isoprenoids > regular steranes > rearranged steranes > tricyclic terpanes (Brooks *et al.*, 1988, Connan, 1984; Seifert and Moldowan, 1979). Therefore we conclude that none of our samples are strongly biodegraded, although they overlap in gravity, stratigraphic and geographic distributions (Bailey *et al.*, 1973b). It is during the early stages of biodegradation, prior to the disappearance of n-alkanes, that biodegradational effects are most likely to mimic maturity variations.

Family A and some Family D oils, probably the equivalent of Williams' (1974) Type I oils, are strongly enriched in n-alkanes compared with branched, cyclic and aromatic compounds (Table 3). This is reflected in the results of factor and correspondence analysis (Snowdon and Osadetz, 1988b). However, the recognition of n-alkane enriched Family D oils from Winnipegosis pools, among more recent discoveries, suggests that it is unlikely that any single oil family can be distinguished using its gasoline range composition alone.

Family A and select Family D n-alkane enriched compositions are also recognizable on cross-plots using isopentane/n-pentane, 3-methylpentane/ nhexane, n-hexane/methylcyclopentane, n-heptane/methylcyclohexane and heptane value ratios (e.g. Fig. 7). Otherwise, there is a considerable overlap in gasoline range compositions, that is consistent with the factor and correspondence analyses. This overlap is illustrated by compositional ratios involving aromatic compounds like, 3-methylpentane/benzene, cyclohexane/ benzene and methylcyclohexane/toluene (e.g. Fig. 8).

Table 5: Oil sample 210°C Saturate Fraction Compositional Characteristics

GSC#	PROD. ZONE	%HC	S/A	P/Ph	23/30	Ts/Tm	D/R	S/R	BB/R	HP
FAMILY A										
0549	Bighorn	63.7	0.64	1.29	0.03	0.61	2.36	1.05	1.07	34
0550	Bighorn	77.0	0.82	1.53	0.03	1.00	1.73	1.25	1.61	34
0920	Bighorn	61.7	0.66	1.11	0.01	0.60	1.93	0.99	1.08	34
1165	Winnipegosis	87.8	1.77	0.55	0.08	0.90	4.57	0.90	0.84	34
6979	Bighorn	70.8	1.60	0.22	0.04	0.71	4.75	1.21	1.21	34
7597	Bighorn	46.1	1.17	1.10	0.06	0.73	5.82	1.44	1.02	34
FAMILY B										
0515	Bakken	90.6	1.16	1.52	0.61	1.17	1.17	0.77	0.91	N
0554	Bakken	93.0	1.47	1.76	0.40	1.65	1.43	1.13	1.06	N
1279	Bakken	91.3	1.48	1.40	0.84	1.18	2.67	0.80	0.76	N
FAMILY C										
0494	Frobisher	78.5	0.66	0.70	0.31	0.41	0.39	0.73	1.04	35
0495	Tilston/Souris	86.8	1.18	0.61	0.45	0.69	0.52	0.97	1.28	35
0497	Forbisher	75.9	0.84	0.64	0.34	0.58	0.29	0.86	1.04	N
0499	Midale	95.0	1.57	0.87	0.69	2.21	1.21	1.59	1.77	N
0500	Midale	79.8	0.76	0.50	0.49	0.61	0.24	0.75	0.94	35
0511	Tilston	79.5	0.84	0.56	0.43	0.66	0.35	0.92	1.16	35
0512	Jurassic	80.9	0.75	0.67	0.40	0.52	0.33	0.98	1.15	35
0514	Midale	80.6	0.69	0.91	0.39	0.32	0.49	0.91	1.38	35
0516	Alida	94.5	1.53	0.87	0.45	1.78	0.99	0.98	1.32	35
0520	Frobisher	89.3	1.26	0.62	0.37	1.00	0.36	0.95	1.34	35
0523	Wapella	82.8	0.87	0.80	0.31	0.47	0.34	0.95	1.26	35
0533	Frobish/Alida	87.5	0.90	0.76	0.48	0.48	0.37	0.91	1.51	35
0539	Alida	82.6	0.63	0.55	0.43	0.55	0.37	0.92	1.06	35
0540	Ratcliffe	86.6	0.75	0.95	0.30	0.55	0.57	0.90	0.92	35
0547	Midale	81.3	1.04	0.51	0.48	0.77	0.31	1.05	1.56	35
0548	Frobish/Alida	86.2	1.12	0.60	0.38	0.83	0.35	0.92	1.30	35
0553	Bakken	87.2	1.04	1.03	0.45	0.33	0.71	0.80	0.87	35
0555	Frobish/Alida	85.9	0.89	0.84	0.41	0.46	0.51	0.95	1.02	35
0556	Ratcliffe	85.2	0.83	0.72	0.44	0.41	0.37	1.09	1.37	35
0557	Ratcliffe	81.0	0.59	0.48	0.36	0.37	0.18	0.86	0.80	N
0559	Ratcliffe	80.6	0.64	0.65	0.37	0.32	0.35	0.76	0.82	35
0560	Watrous	82.8	0.84	0.52	0.48	0.47	0.33	1.03	0.97	35
0565	Alida/Tilston	86.8	1.25	0.77	0.39	1.19	0.61	1.15	1.38	35
0585	Alida	80.0	0.68	0.58	0.31	0.43	0.36	0.80	1.12	35
0595	SourisValley	86.7	1.43	0.77	0.39	0.67	0.34	0.86	1.37	35
0596	Tilston	85.1	0.90	0.65	0.31	0.54	0.32	1.04	1.02	35
0711	MissCanyon1B	78.9	1.46	0.78	0.70	1.57	1.02	0.79	0.76	35
0714	MissCanyon3B	88.1	1.32	0.75	0.54	1.29	1.04	0.79	0.88	35
0717	Lwr.Amaranth	90.8	1.55	1.04	0.68	1.68	0.96	0.81	0.94	35
0721	LodgepoleA	85.5	1.09	0.81	0.90	0.74	0.35	0.76	0.93	35
0722	LodgepoleB	88.0	1.19	0.72	0.65	1.04	0.44	0.72	0.78	35
0725	LodgepoleA	88.3	1.15	0.62	0.75	0.94	0.48	0.64	0.63	35
FAMILY D										
0466	Birdbear	91.1	1.49	1.47	0.04	0.47	1.21	0.59	0.74	34
0546	Ratcliffe	36.5	0.81	0.86	0.25	0.34	0.51	0.88	1.38	34
0551	Birdbear	90.7	1.37	1.20	0.08	0.44	1.51	1.01	1.08	34
0552	Birdbear	91.1	1.32	1.17	0.07	0.48	1.14	0.94	1.17	34
0558	Ratcliffe	87.0	1.19	0.97	0.19	0.94	1.53	0.93	1.62	34
0568	Ratcliffe	81.4	0.97	0.75	0.09	0.97	1.56	1.07	0.80	34
0577	Ratcliffe	65.8	1.59	1.05	0.14	0.94	1.90	1.10	0.83	34
0755	Winnipegosis	89.5	2.59	0.62	0.08	1.12	1.58	1.17	1.34	34
0756	Winnipegosis	93.6	2.51	0.75	0.09	1.16	1.73	1.09	1.30	34
0800	Winnipegosis	91.4	1.89	0.82	0.06	1.09	3.17	1.34	1.19	34
0801	Winnipegosis	90.1	1.57	0.82	0.09	1.03	4.00	1.08	1.17	34
0802	Winnipegosis	86.3	1.51	0.83	0.12	0.69	0.95	0.78	0.73	34
0841	Winnipegosis	59.6	1.02	0.81	0.04	0.42	0.57	0.74	0.81	34
0842	Winnipegosis	71.1	1.10	0.79	0.03	0.30	0.69	0.77	0.60	34
0924	Winnipegosis	89.7	1.77	0.68	0.05	0.96	1.81	1.09	0.98	34
1275	Winnipegosis	92.1	2.44	0.88	0.06	1.00	1.85	1.19	0.94	34
1276	Winnipegosis	92.7	2.73	0.76	0.05	1.00	2.08	1.19	1.26	34

Legend to Table 5:

%HC = percentage of hydrocarbons in 210° C fraction

S/A = saturate hydrocarbon/aromatic hydrocarbon ratio in 210° C fraction

P/Ph = pristane/phytane peak area ratio from 210° C fraction saturate fraction gas chromatogram

23/30 = a/f = C23 tricyclic terpane/17alpha(H)-hopane, C30 pentacyclic terpane

Ts/Tm = c/d = Ts [18alpha(H)-trisnorhopane]/Tm [17alpha(H)- trisnorhopane]

D/R = 6/14 = 13beta(H), 17alpha(H)-diacholestane (20S)/5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20R)

S/R = 20/23 = 24-ethyl-5alpha(H), 14alpha(H), 17alpha(H)- cholestane (20S)/24-ethyl-5alpha(H), 14alpha(H), 17alpha(H)- cholestane (20R)

BB/R = 21/23 = 24-ethyl-5alpha(H), 14beta(H), 17beta(H)- cholestane (20R)/24-ethyl-5alpha(H), 14alpha(H), 17alpha(H)- cholestane (20R)

HP = carbon number prominence amongst extended hopanes, peaks h- q, on m/z 191 mass chromatograms; N - no prominence; 35 - pentakishomohopanes predominant; 34 - tetrakishomohopanes predominant.

Table 6: Saturate fraction biological marker compounds identified as peaks on m/z 217 sterane traces

2 — C21 sterane
3 — C21 sterane
5 — C22 sterane
6 — 13beta(H), 17alpha(H)-diacholestane (20S)
7 — 13beta(H), 17alpha(H)-diacholestane (20R)
10 — 5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20S)
11 — 5alpha(H), 14beta(H), 17beta(H)-cholestane (20R)
12 — 24-ethyl-13beta(H), 17alpha(H)-diacholestane (20S)
13 — 5alpha(H), 14beta(H), 17beta(H)-cholestane (20S)
14 — 5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20R)
15 — 24-ethyl-13beta(H), 17alpha(H)-diacholestane (20R)
16 — 24-methyl-5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20S)
17 — 24-methyl-5alpha(H), 14beta(H), 17beta(H)-cholestane (20R)
18 — 24-methyl-5alpha(H), 14beta(H), 17beta(H)-cholestane (20S)
19 — 24-methyl-5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20R)
20 — 24-ethyl-5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20S)
21 — 24-ethyl-5alpha(H), 14beta(H), 17beta(H)-cholestane (20R)
22 — 24-ethyl-5alpha(H), 14beta(H), 17beta(H)-cholestane (20S)
23 — 24-ethyl-5alpha(H), 14alpha(H), 17alpha(H)-cholestane (20R)

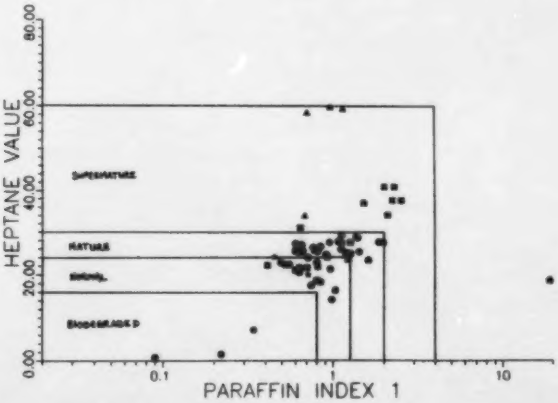

Figure 7: Cross-plot of Paraffin Index 1 versus Heptane Value showing boundaries of biodegradation and thermal maturity classes as inferred by Thompson (1983). Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares. Thermal maturity classes suggested by Thompson (1983) are not considered indicative of thermal maturity when this figure is compared with figure 11.

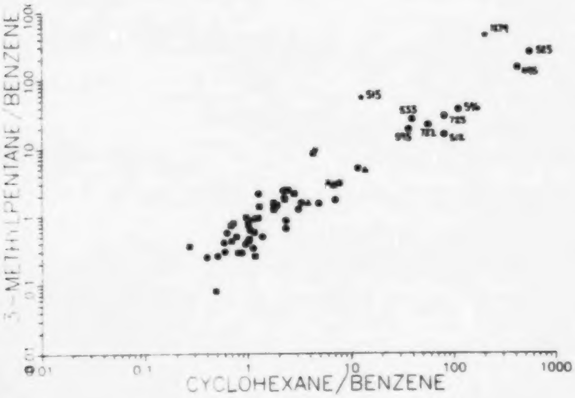

Figure 8: Cross-plot of cyclohexane/benzene versus 3-methylpentane/benzene ratio. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares. Both these ratios are commonly inferred to indicate water washing. Note that some of the most strongly water washed oils, particularly Family B oils, are amongst the highest API gravity oils.

ORIGIN AND SIGNIFICANCE OF SELECT COMPOSITIONAL VARIATIONS

THERMAL MATURITY

For some compositional characteristics sensitivities to maturity have been established and these are used primarily as thermal maturity indicators (Waples and Machihara, 1990; Curiale *et al.*, 1989; Philp, 1985; Mackenzie, 1984; Thompson, 1983). Sterane thermal maturity indicators are generally preferred, but because their epimerization reactions are generally complete during early significant hydrocarbon generation (*ibid.*) and because of limits on their thermal stablility (Osadetz *et al.*, in review) they cannot adequately describe progressive maturity differences amongst oils. The trisnorhopane maturity ratio (Ts/Tm) (Fig. 2) has the potential to describe oil pool thermal maturity. The relationship proposed by Osadetz *et al.* (in review), that is discussed below, appears consistent with other constraints on Williston Basin oil pool maturity. Gasoline range thermal maturity (Fig. 7) indicators (Thompson, 1983) would be desirable because of their great abundance and high thermal stabilities, however, as discussed below, they are of limited value.

Williston Basin Ts/Tm variations consistently characterize oil pool maturities for all families and independently match critical hydrocarbon generation and expulsion thresholds defined by Rock-Eval/TOC and extract gross compositions. This matching is facilitated by model vitrinite reflectance (VR) values, inferred from the relationship between vitrinite reflectance and the C_{29} sterane S/R thermal maturity ratio (attributed to Zumberge, by Bein and Sofer, 1987) as substituted into the linear relationship between Ts/Tm and C_{29} sterane S/R ratio defined empirically from Family D oils (Fig. 9).

The resulting relationship is:

$$VR = 0.245 \times (Ts/Tm) + 0.626$$

Many Family A and C oils also conform to this equation, but other compositional variations suggest some steranes are thermally degraded above VR = 0.9 percent (Fig. 9). Samples with high Ts/Tm, > 0.66, and other compositional variations suggesting thermal cracking have lower than expected C_{29} S/R sterane ratios.

The Ts/Tm-VR correlation is supported by maturity variations in oil families and is inferred to be source rock independent. The lowest Ts/Tm ratios in both Family C (#514 and # 559, Ts/Tm = 0.32) and Family D (#842, Ts/Tm = .30) suggest VR = 0.70, consistent with an intense hydrocarbon generation threshold for Type II marine source rocks (Powell and Snowdon, 1983), but very different than the VR = 1.1 percent inferred from gasoline range hydrocarbon (Thompson, 1983). The least mature Family B oil (#515, Ts/Tm = 1.17) has a VR = 0.91 percent, consistent with a Bakken Formation expulsion threshold inferred from extract and Rock-Eval data (Transformation Ratio = 68 percent at approximately VR = 0.90 percent; Osadetz and Snowdon, in review). The most immature Family A composition (#920, Ts/Tm = 0.60) indicates maturities equivalent to VR = 0.77 percent, matching the main hydrocarbon generation stage and expulsion threshold inferrred for Ordovician kukersites.

This internal agreement among oil pool compositions is unusual because facies sensitivity has often frustrated Ts/Tm application to maturity characterizations (Waples and Machihara, 1990). Source rock extracts do not exhibit the same consistent Ts/Tm variations, nor do source rock molecular maturity indicators agree with extract %HC criteria for thermal maturity (Osadetz *et al.*, in review). That oil pool compositions vary so systematically, while extracts behave so poorly, suggests either very specific facies source oil pools or that processes affect molecular species partition in expelled and migrated hydrocarbons.

Thompson (1983; 1979) proposed that several gasoline range compositional ratios, including, Heptane value, Paraffin Index 1, n-heptane/ methylcyclohexane, and n-hexane/ methylcyclopentane, were maturity dependent. He suggested that specific Paraffin Index 1 and Heptane values could be used to distinguish biodegraded oils from progressively more mature oils. Williston Basin oils do show a general positive correlation between these two ratios but not with other maturity indications (Fig. 7). When gasoline range maturity indicators are cross-plotted with Ts/Tm ratio there is no consistent variation (Fig. 11), although Family D oils do exhibit a slight increase in Paraffin Index 1 and n-hexane/methylcyclopentane ratio with increasing Ts/Tm. Neither do gasoline range thermal maturity ratios show systematic variation with API gravity or S/A HC ratio. Such cross-plots indicate that Family A and some Family B oils have consistently high gasoline range ratios, because of their enrichment in n- alkanes, while three Family C oils, Moose Valley Tilston (#596), Wapella Mannville (#523) and Red Jacket Shaunavon (#512), have consistently low values (Fig. 11). These last three samples have low API gravities compared to oils of comparable Ts/Tm thermal maturity and all three are interpreted to be biodegraded.

Progressive or systematic changes in Family C S/A HC ratio, DIA/REG sterane ratio and API gravity all follow Ts/Tm thermal maturity. Samples #511 and #595 define the limits between two Family C compositional groups (Fig. 12). Oils less mature than #511 have DIA/REG < 0.6 and S/A HC < 1.0. When Ts/Tm reaches 0.67, S/A HC exceeds 1, while DIA/REG remains below 0.6, although Family C SFGC and terpane compositional characteristics persist. When Ts/Tm exceed 1.1, DIA/REG exceeds 0.6 and increases with increasing Ts/Tm to 2.1 accompanied by a slight increase in S/A HC. Family C oils with Ts/Tm > 1.1 and DIA/REG 0.6 also have Pr/Ph > 0.75, but not exceeding 1.1. The relationship between API gravity and Ts/Tm is also strong and oils > 33° API generally have Ts/Tm greater than 0.65 and commonly have S/A HC ratio > 1 (Fig. 13).

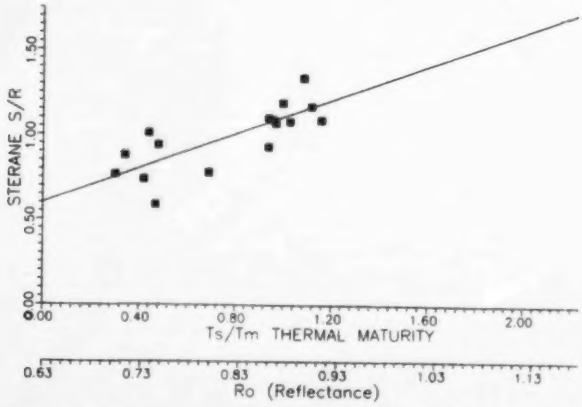

Figure 9: Oil Family D trisnorhopane (Ts/Tm) maturity ratio versus C29 sterane S/R maturity ratio cross-plot. The regression line, S/R = 0.503 (Ts/Tm) + 0.602, suggests a relationship between vitrinite reflectance, VR = 0.245 (Ts/Tm) + 0.626, following a correlation between VR and C29 sterane S/R that Bein and Sofer (1987) attribute to Zumberge.

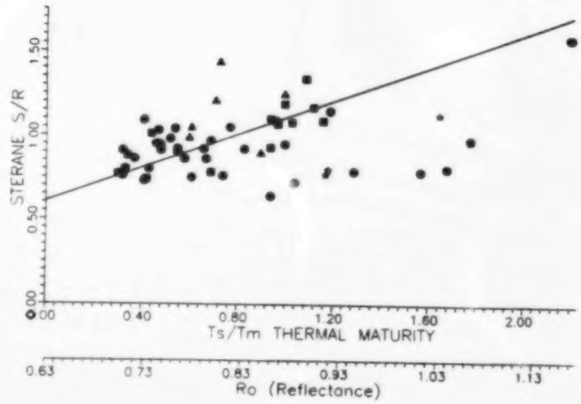

Figure 10: Oil pool trisnorhopane (Ts/Tm) maturity ratio versus C29 sterane S/R maturity ratio cross-plot. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares. The regression line and relationship with vitrinite reflectance is like that in figure 30. Notice how more mature (Ts/Tm and API gravity) Family B and C oils have anomalously low S/R ratios.

Osadetz *et al.* (in review) observed a geographic significance to Family C thermal maturity variations. High Ts/Tm, S/A HC ratios and API gravities occurred southwest of a line trending northeast between Steelman (#499) and Parkman (#595) fields on the one side and Kenosee field (#511) on the other. From this they concluded that Family C oil pool thermal maturity decreases northwestwardly along the Madison subcrop. No consistent stratigraphical control on thermal maturity could be established, although they did suggest that Family C oils in the Oungre trend (Hartling *et al.*, 1982) were locally sourced.

Osadetz *et al.* (in review) inferred that regular steranes become thermally unstable at Ts/Tm > 1.1 (approximately VR = 0.9 percent). Their cracking increases DIA/REG in more mature oils. This precedes cracking of less complex aliphatic compounds at VR = 1.2 percent, the classical transition zone. All Family B oils have DIA/REG > 1, thermal maturities greater than VR = 0.9 percent and oil gravities only slightly higher than Family C oils of comparable thermal maturity (Fig. 13). Because the Bakken expulsion threshold coincides with regular sterane thermal stability, it is not surprising that Bakken oil pools do not reflect the wider extract DIA/REG ratios and gravities observed in Family C oils at lower maturities. Therefore the quality of Bakken oils is controlled directly by a lithologically dependent expulsion threshold. This effect reduces the size of the region where Bakken sources are sufficently mature to expel oils but it makes probable that Bakken oils will be of high gravity.

Family D oils show the strongest dependence of S/A HC ratio and API gravity on Ts/Tm thermal maturity (Figs. 12, 13). Both increase progressively and systematically with increasing maturity and the linear nature is more pronounced if only Winnipeogsis reef pools are considered. Assuming that Ts/Tm ratios are completely unaffected by biodegradation, it is clear that thermal maturity exerts a dominant control on S/A HC ratio and API gravity.

There are relatively few Family A oils and associations amongst Ts/Tm thermal maturity, nor are S/A HC ratio and API gravity well established (Figs. 12, 13). There is a strong tendency for the denser oils (#920, #549) to have significantly lower Ts/Tm (approximately 0.6) compared with the higher gravity oils (#550, #1165) whose Ts/Tm values are between 0.9 and 1.0.

BIODEGRADATION

From the susceptibility of n-alkanes to biodegradation and the presence of significant n-alkanes in both the gasoline range and > 210°C saturate fraction gas chromatograms, we conclude that biodegradation of our sample set is slight if present at all. Lower molecular weight n-alkanes should be more susceptible and their depletion should be reflected in the gasoline range first. High isopentane/n-pentane and 3-methylpentane/n-hexane ratios (Fig. 14) are commonly used biodegradation indicators, but absolute criteria describing degree of biodegradation have not been defined. Initial differences in the concentration of n-alkanes to branched and cyclic compounds that are attributable to source rocks composition, as is the case for Ordovician oils, may result in similar gasoline range ratios reflecting differing amounts of biodegradation. Added to this are competing maturity effects that tend to increase n-alkane concentration with increasing thermal maturity (Thompson, 1983). These render the identification of biodegradational effects difficult. However, we shall show that it is possible to recognize biodegradation as a departure from normal maturation variations attributable to gasoline range n-alkane depletion.

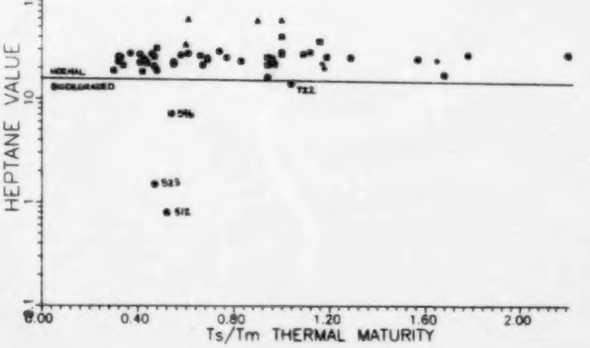

Figure 11: Cross plot of Ts/Tm versus Heptane Value indicating the lack of variation in Heptane Value with inferred thermal maturity and anomalously low Heptane Values of samples #512, #523, and #596. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares.

Samples show a positive correlation between isopentane/n-pentane and 3methylpentane/n-hexane ratios with a few outliers (Fig. 14). Not surprisingly, the lowest values are shown by n-alkane enriched Family C and Family D oils, thereby indicating that source characteristics play an important role in controlling these ratios. However, there are two Family C samples from Wapella and Red Jacket (#512 and #523), with the highest 3methylpentane/n-hexane ratios that consistently fall outside the thermal maturity variation exhibited by other oils on cross plots of Ts/Tm versus Heptane Value (Fig. 11), Paraffin Index 1, n-hexane/methylcyclopentane, nheptane/methylcyclohexane, 3-methylpentante/n-hexane and fall into the biodegraded field of the Paraffin Index 1 versus Heptane Value cross-plot (Fig. 7).

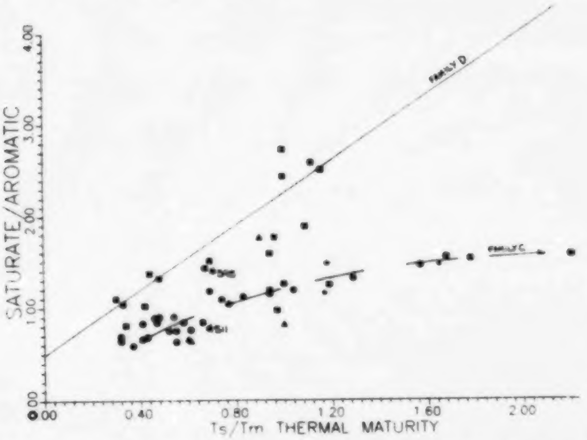

Figure 12: Cross plot of Ts/Tm versus saturate/aromatic hydrocarbon ratio. Notice the progressive but different changes in Family C and Family D saturate/aromatic ratio with increasing thermal maturity. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares.

Both oils lie at the northern and northwestern limits of the sample region and had high gasoline range Factor 3 (biodegradation) scores (Snowdon and Osadetz, 1988b). Another Family C sample, Moose Valley Tilston (#596), is also depleted of n-hexane and n-heptane. This sample is geographically close to Red Jacket (#512) and Wapella (#523), has a high gasoline range Factor 3 score (ibid.) and follows the other two samples on several of the cross plots mentioned above. Yet it remains within the standard variation of unbiodegraded oils on cross plots involving Ts/Tm, isopentane/n-pentane and 3methylpentane/n-hexane (Fig. 11). This pool is also interpreted to be biodegraded, although less so than Wapella and Red Jacket. Other Family C oils occurring near the western limit of established production (#494, #514, #585, #557) lack high 3-methylpentane/n-hexane and isopentane/n-pentane ratios, are not depleted in either n-hexane or n-heptane, and conform to other gasoline range variations consistent with thermal maturity.

3-methylpentane/n-hexane ratio decreases with increasing Ts/Tm (Fig. 15). Family A oils, from Type I source rocks behave differently from the other three families which are derived from Type II source rocks. Even amongst the Type II sourced oils the correlations are slightly different and 3methylpentane/n-hexane decreases most rapidly with increasing maturity for Family D oils. S/A HC ratio and API gravity are also inversely correlated with 3-methylpentane/n-hexane ratio, but because of the correlation between maturity and 3-methylpentane/n-hexane, we conclude that these variations are also dominated by thermal maturity.

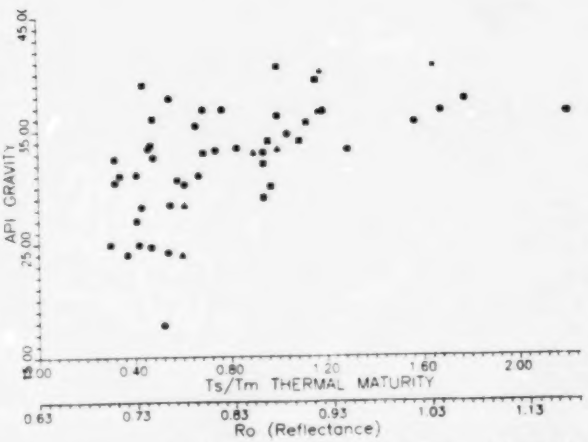

Figure 13: Cross plot of Ts/Tm versus API gravity. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares.

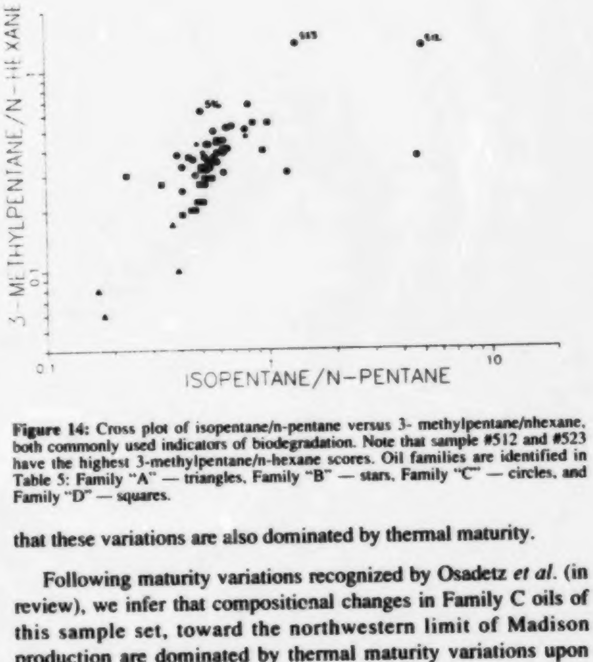

Figure 14: Cross plot of isopentane/n-pentane versus 3- methylpentane/nhexane, both commonly used indicators of biodegradation. Note that sample #512 and #523 have the highest 3-methylpentane/n-hexane scores. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares.

Following maturity variations recognized by Osadetz *et al.* (in review), we infer that compositional changes in Family C oils of this sample set, toward the northwestern limit of Madison production are dominated by thermal maturity variations upon which biodegradation has been superimposed in some instances. Other oil families show no evidence of any biodegradation.

WATER WASHING

Water washing is commonly recognized by depletion of the more soluble compounds, particularly benzene and toluene (McAuliffe, 1978). The effects of water washing should be progressive, with the loss of more soluble compounds always exceeding that of the less soluble. Despite the low solublities of heavier n-alkanes, losses of compounds up to nC15 have been attributed to water washing (Hunt, 1979, p. 390). Ratios of cyclohexane/benzene, 3methylpentane/benzene and methylcyclohexane/toluene are commonly used indicators of water washing.

Snowdon and Osadetz (1988b) found that gasoline range Factor 2 gave high negative loadings to benzene and toluene. This factor carried a more significant proportion of the variance than gasoline range Factor 3, which was partly attributed to biodegradation, thereby suggesting that water washing had a more important effect than biodegration. Logarithmic cross plots of cyclohexane/benzene, 3-methylpentane/benzene and methylcyclohexane/toluene show positive correlations amongst each other, with the strongest being the correlation of logarithm cyclohexane/benzene with logarithm 3methylpentane/ benzene (Fig. 8).

Each oil family exhibits different variations of water washing parameters, but all fall on a similar trend suggesting that difference in source has little effect on these parameters. Some of the highest gravity and most mature oils are amongst the most strongly water washed. There is no consistent correlation of S/A HC ratio or API gravity and water washing alone.

The three recognizably biodegraded oils are substantially water washed but water washed oils are generally not biodegraded and we conclude that connate water washing predominated over meteoric water washing. Samples with some of the highest cyclohexane/benzene and 3-methylpentane/benzene ratios (#515, #523, #1279, #495, #597) occur near the geographical limits of production and are presumably amongst the longest migrated oils. As for Family D oils, those with the highest cyclohexane/benzene and 3-methylpentane/benzene ratios, include some of the most mature oils that now occur far up dip of their site of generation. Both of these associations suggest that extensive connate water washing is sometimes associated with the length of secondary migration pathways.

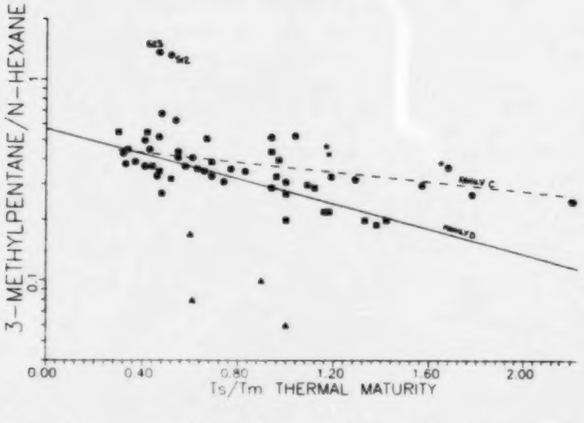

Figure 15: Cross plot of Ts/Tm versus 3-methylpentane/n-hexane. The regression lines are to Family D (solid) and Family C (dashed) data. Notice how biodegraded samples #512 and #523 stand out from the thermal maturity dominated variations reflected by the other samples. Oil families are identified in Table 5: Family "A" — triangles, Family "B" — stars, Family "C" — circles, and Family "D" — squares.

SUMMARY

Source rock composition and thermal maturity are important controls on Ts/Tm ratio, API gravity, S/A ratio, and certain gasoline range compositional characteristics. Each family of oils responds differently to thermal maturation. Family C and D oils have standard Type II expulsion thresholds (VR = 0.7 percent), but variation of oil gravity and S/A ratio in Family D oils is much greater than that observed in Family C. Family B and A oils have different expulsion thresholds due to variations in source rock lithology and organic matter type. Family A oils have Type I source rocks that enter the main hydrocarbon generation stage at VR = 0.77 percent and expel their oils immediately. Like Family D, Family A oil gravities appear to depend significantly on thermal maturity. Bakken source rocks, on the other hand, enter the oil window at VR = 0.7 percent, like other Type II source rocks, but expulsion is delayed until approximately VR = 0.9 percent. This has a significant effect on oil gravity and S/A HC ratio as well as other compositional characteristics. Unaltered Family B oils will have consistently high gravities because of this.

The effects of early biodegradation are difficult to separate from low thermal maturity. Only three biodegraded Family C oils could be clearly identified (#512, #523, #596). Other oil families show no signs of biodegradation. Water washing is more extensive than previously described and it affects all oil families to differing degrees. The lack of associated biodegradation but the strong association with oil pool maturity and migration pathway lengths suggest that connate water washing predominates and that this may occur during secondary migration as well as following accumulation.

RELATIONSHIPS BETWEEN OIL QUALITY AND MIGRATION PATHWAY

Knowing that oil quality and oil pool maturity are related, we can compare oil pool maturity and geographic location with source rock maturity patterns to characterize migration pathways. Source rock maturity - depth relationships can be described using model vitrinite reflectance profiles, that are a function of thermal history. Source rock extracts, Rock-Eval pyrolysis, Bakken vitrinite reflectance and kinetic maturation models (Osadetz and Snowdon, in review; Gosnold and Huang, 1987; Webster, 1984; Price *et al.*, 1984), all suggest geographically variable, time-invarient, effective heat flows generally less than 67 mW/m^2. Regions outside the enhanced hydrocarbon generation domain appear to have time-invarient, effective heat flows of 50 to 60 mW/m^2, although even lower effective heat flows are suggested by the deepest marginally mature Bakken source rocks. By combining source rock maturity and depth (for specific thermal histories) with API gravity as a function of oil pool maturity, source rock evolution lines can be plotted that describe the API gravity that a source rock would expel. The gravities of oils that occur at depths shallower than that predicted by the appropriate source rock oil gravity evolution lines, quantitatively display vertical migration. Where oil pools occur in the same stratigraphic interval as their source rock, as is the general case in Williston Basin, the vertical migration distance will be related to lateral migration distance through the dip. This analysis allows us to characterize oil family migration pathways.

API GRAVITY THERMAL MATURITY RELATIONSHIPS

Family D and Family C oils have enough members to describe the variation of API gravity with maturity. The Family D Ts/Tm versus API relationship is linear (Fig. 16) but an exponential relationship best fits the Family C data (Fig. 17). Family D oils exhibit a limited range of maturities, generally less than VR = 0.9 percent, whereas some Family C oils are much more mature. The Family C Ts/Tm versus API gravity relationship probably involves some thermal cracking, as indicated by Ts/Tm versus DIA/REG variations, but most differences in the Family D and Family C Ts/Tm versus API gravity relationships, reflect real differences in the density of oils arising from the two source rocks. Family D variations suggest that a wide range of API gravities, between 25° and 50° will occur within the oil window (0.7 percent < VR < 1.2 percent) and that there will be a strong, progressive increase in API gravity with increasing thermal maturity. In contrast, Family C variations indicate a more limited range of API gravities, between approximately 25° and 40°, with considerable variation in gravity below VR = 0.8 percent and little variation at higher maturities. Gasoline range data indicate that many samples are affected by water washing and that three Family C oils are affected by biodegradation, but these processes have minor effects on API gravity and do not detract from the foregoing inferences.

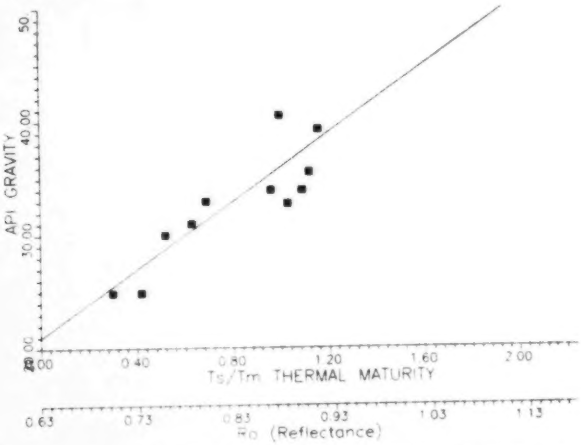

Figure 16: Winnipegosis reef oil pool, Family D, trisnorhopane (Ts/Tm) maturity ratio versus API gravity cross-plot. The regression line, API = 14.99 (Ts/Tm) + 20.98, indicates that thermal maturity is a primary control on API gravity. Not all data used to form the regression line are shown. The relationship between Ts/Tm ratio and VR inferred from Fig. 9 is also shown.

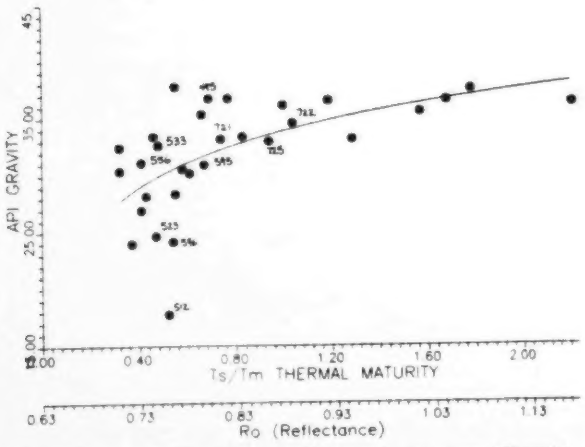

Figure 17: Family C oils, trisnorhopane (Ts/Tm) maturity ratio versus API gravity cross-plot. The regression curve, indicates a limited range of oil gravities that are less sensitive to thermal maturity than Family D oils. Ten samples whose gasoline range compositions suggest either water washing or biodegradation are identified. With the exception of Red Jacket (#512), Wapella (#523) and Moose Valley (#596) pools biodegradation and water washing appear to have little effect on API gravity, when compared with unaffected samples. Note the low gravity of Neptune Ratcliffe pool (#557) that cannot be attributed to either water washing or biodegradation. The relationship between Ts/Tm ratio and VR inferred from Fig. 9 is also shown.

FAMILY D — WINNIPEGOSIS MIGRATION CHARACTERISTICS

Family D oils plotted on the API gravity vs. depth diagram show a wide scatter (Fig. 18). If pool positions are corrected for stratigraphic separation, reservoir and source rock associations are clearer. Several pools are located on the margin of the enhanced hydrocarbon generation region. These include Oxbow Winnipegosis pool (#841 and #842), Hoffer Ratcliffe pools (#558, #568), Flat Lake Ratcliffe pool (#577) and Freda Lake Ratcliffe pool (#546). All have maturities similar to Winnipegosis sources underlying the pools. This suggests local sourcing, although Ratcliffe Beds pools have an obvious cross-stratal migration component. Ratcliffe Beds Family D oil pools overlie the eastern Winnipegosis Platform margin and one may speculate that they are sourced by platform rather than basin sources.

Some Winnipegosis reef oils have maturities suggestive of local sources. These include Bienfait (#802), Macoun 16-11 (#801), and several small pinnacles in the Tableland area (#755, #924, #800). Oils in large reef build-ups in both the Tableland (#756) and Macoun (#1275; #1276) areas have significantly higher maturities than nearby source rocks and therefore must have migrated from deeper in the basin. The most obvious migration conduit is the Lower Member platform carbonates.

Birdbear pools at Wapole (#466) and Kisbey (#552) indicate long migration pathways, both vertically and laterally. Lateral migration components are > 3200 m. Only 450 m can be attributed to the vertical stratigraphic separation between reservoir and source rock. Both oils could have been sourced in Canadian parts of the basin, however it is unknown whether the lateral migration was restricted to the reservoir horizon or else, included the more distant Lower Winnipegosis platform. Local dislocation from salt dissolution at Kisbey might lend credence to the latter suggestion.

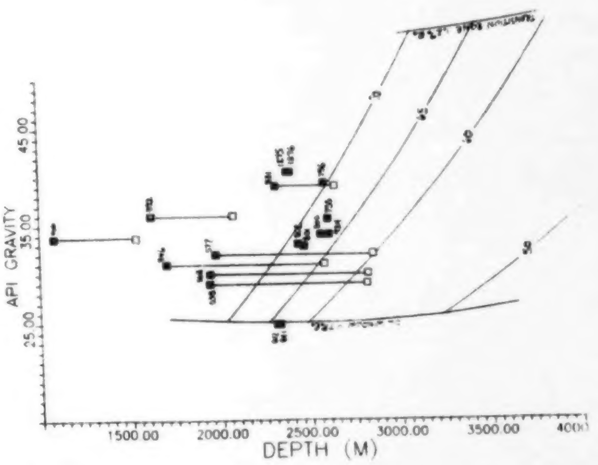

Figure 18: Oil gravity versus depth cross-plot for Family D oils. The contruction of source rock API gravity-depth relationships for various steady state effective heat flows are discussed in the text. The oil window (VR = 0.7 percent) and transition zone (VR = 1.2 percent) boundaries are shown. Where Family D oils occur in stratigraphic horizons other than the Winnipegosis Formation the position of underlying Winnipegosis Formation is indicated by the associated open square. In most cases oil gravities match those expected from geographically nearby source rocks indicating that most Family D oils have short lateral migration pathways, even if, like the oils in Oungre trend Madison Group reservoirs (#558, #568, #577), they have significant vertical migration pathways. There are several significant exceptions in both Winnipegosis (#756, #1275, #1276), Birdbear (#466, #551, #552) and Madison (#546) reservoirs. These pools must have a significant lateral migration component. The location and lack of lateral migration inferred for Family D Madison Group pools (#558, #568, #577) strongly suggests that they are derived from Winnipegosis platform source rocks.

FAMILY C — LODGEPOLE MIGRATION CHARACTERISTICS

A plot for Family C oils also indicates both distant and locally sourced oils (Fig. 19). Oils in the Oungre area are amongst the deepest of Mississippian production in Canadian Williston Basin.

These are commonly medium gravity oils matching local source maturities. The remaining Family C pools have various gravities occurring over a considerable range of depths. Higher gravity oils, generally > 33° API, tend to occur in more easterly regions, particularly Manitoba. None are strongly biodegraded and lighter gravity oils have compositions consistent with higher maturities. More mature oils tend to occur southeast of the Steelman and Kenosee fields, regardless of stratigraphic position.

Family C pools lying to the northwest tend to have gravities similar to those in the Oungre trend. Biodegradation and water washing do not appear to significantly affect API gravities in most samples (see figure caption for exceptions). While the complete significance of this geographic compositional segregation has not been ascertained, it can be surmised that migrated oils are distinguished by maturity differences reflecting either different lateral migration pathways from a single source or different effective source beds in the Lodgepole.

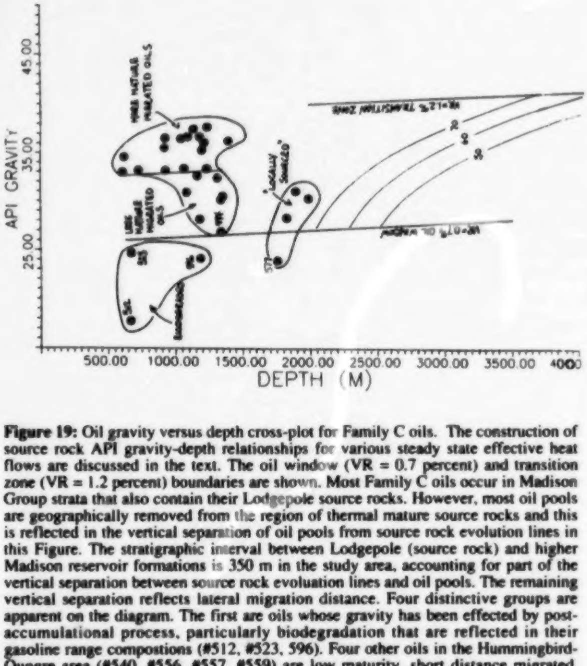

Figure 19: Oil gravity versus depth cross-plot for Family C oils. The construction of source rock API gravity-depth relationships for various steady state effective heat flows are discussed in the text. The oil window (VR = 0.7 percent) and transition zone (VR = 1.2 percent) boundaries are shown. Most Family C oils occur in Madison Group strata that also contain their Lodgepole source rocks. However, most oil pools are geographically removed from the region of thermal mature source rocks and this is reflected in the vertical separation of oil pools from source rock evolution lines in this Figure. The stratigraphic interval between Lodgepole (source rock) and higher Madison reservoir formations is 350 m in the study area, accounting for part of the vertical separation between source rock evolution lines and oil pools. The remaining vertical separation reflects lateral migration distance. Four distinctive groups are apparent on the diagram. The first are oils whose gravity has been effected by post-accumulational process, particularly biodegradation that are reflected in their gasoline range compostions (#512, #523, 596). Four other oils in the Hummingbird-Oungre area (#540, #556, #557, #559) are low maturity, short distance migrated, medium crudes occurring at deeper depths. The remaining oils are long-distance migrated medium and light crudes occurring at shallower depths whose gravities are relatively unaffected by post-accumulation processes. This group can be subdivided into a group of lower maturity, lower gravity < 33° API oils tending to occur in western portions of the Madison Subcrop Petroleum Province and commonly higher maturity (Ts/Tm > 0.66), higher gravity oils (> 33° API) that tend to occur in more easterly positions.

FAMILY A AND B MIGRATION CHARACTERISTICS

Families A and B have either too few samples or too limited a range of thermal maturities to describe meaningful Ts/Tm versus API gravity variations. In this case inferred oil pool maturities and depths can be compared with comparably mature source rocks using model reflectance profiles.

Except for the Minton Winnipegosis pool (#1165), the oils of these two families are enclosed in the stratigraphic unit that contains their source, such that the vertical separation between oil pools and comparably mature source rocks on reflectance depth cross-plots reflect the length of lateral migration pathways. Family B oil pools are generally well removed from mature source regions and occur up-dip of their expulsion thresholds. The Rocanville Bakken pool (#515) occurs more than 200 km away from the region where it was generated and expelled. This observation is consistent with the vertical separation between Family B oils and Bakken source rocks of appropriate maturity. Surprisingly, the very mature Family B pools are sourced by one of the stratigraphically highest source rocks. No oils with inferred maturities > 1.2 percent VR, i.e., the beginning of the transition zone (Powell and Snowdon, 1983), occur in the Canadian portion of the Williston Basin.

The immature, heavy Family A crude in Weirhill Bighorn pool is only slightly removed from both thermal maturity and rich source occurrence limits, and other Family A pools are essentially locally sourced. More mature Family A oils have not been discovered in Canada. Whether or not this is a sampling problem, i.e., that more mature older Paleozoic oils remain to be found in Canada, or that it reflects fundamental limits on the lateral migration of Ordovician oils remains to be proved. However, the long lateral migration distance of Bakken oils (#515, #554, #1279), the better hydraulic characteristics of Ordovician strata when compared with the Bakken Middle member and the few pools in older Paleozoic reservoirs, suggest that the absence of more mature Ordovician sourced oils does not reflect a thermal maturity limitation on Family A potential.

NEW EXPLORATION OPPORTUNITIES

Oil pool accumulation reflects the interplay of source rock distribution, thermal maturation pattern and migration pathways. Differences in reserves are in part due to these factors, but source rock models and end-point migration models suggest significant undiscovered potential.

The maturity versus depth cross-plot indicates that Ordovician sourced oils are characterized by short migration pathways, but Bakken sourced oils are characterized by very long migration pathways. Thus the full range of Ordovician source rock maturities are not represented by current Canadian discoveries. Bighorn sources have fed the accumulations in the Cedar Creek Anticline and recent discoveries at Minton and Weirhill in Canada indicate the potential for radial migration pathways into the northeastern and northwestern parts of the Williston Basin. The difference in structural history between the Nesson and Cedar Creek anticlines and the timing of hydrocarbon generation in Ordovician sources deduced from model studies, suggest a significant, untested, conceptual play at shallow depths in Canada and North Dakota. Primary or diagenetic porosity associated with an inferred Upper Ordovician Yeoman shoal-trend bordering a restricted platform facies belt (Kendall, 1976), could have entrapped Bighorn Group oils that migrated up-dip before the Nesson Anticline became a significant barrier.

The few pools from Bakken source rocks indicate complex migration pathways of great length. Rocanville and Daly Bakken oil pools occur in sparsely explored, uppermost Devonian deposits. These shoreline sands may provide a suitable migration conduit for Bakken oils migrated from the United States. Complex Madison Group stratification patterns, so apparent in its control on source rock distribution, offer new and interesting possibilities for internal stratigraphic and diagenetic traps for Family C oils.

Winnipegosis oil-source systems indicate significant potential in both shelf edge and pinnacle reef plays. The former have been well explored in the United States and the latter have been well established in Canada with discovery of more than fifteen new pools. There is clear potential in the Canadian platform margin and in American pinnacle reefs. This study shows that long lateral migration pathways through the Lower Winnipegosis Platform exist and that reefs may be prospective up dip of the region of thermally mature source rocks. The best Winnipegosis reef prospects will be either large coalesced pinnacle build-ups in the basin or platform margin plays connected to regional migration pathways. Using the API gravity versus depth diagram, and the amounts of vertical separation between oils and comparably mature source rocks we infer that lateral migration pathways up to 50 km beyond the oil window can be expected. The delineation of these migration pathways will be difficult, but possibilities of concentrated large reserves at reduced drilling depths render the search for them attractive.

Finally, source rock distributions and maturity patterns indicate that the most favourable exploration regions occur within the region of established production. In western Saskatchewan increased maximum burial depth may well have compensated for decreased crustal heat flows and a new oil window established away from the region of enhanced hydrocarbon generation. Through central Sakatchewan, long migration paths, insufficient thermal maturity, lack of older Paleozoic source rock, and biodegradation accompanying fresh formation water recharge, all increase the risk associated with exploring in that region.

CONCLUSIONS

Four compositionally distinctive oil families occur in Canadian Williston Basin. Generation and expulsion thresholds vary with kerogen composition and thermal history. Ordovician Type I sources expel oils at VR = 0.77 percent accompanying significant hydrocarbon generation. Winnipegosis and Lodgepole Type II sources also expel oils at the start of significant hydrocarbon generation, which they enter at VR = 0.70 percent. Bakken sources enter the main hydrocarbon generation stage at VR = 0.70 percent, like other Type II sources, but expulsion is delayed until higher maturities around VR = 0.90 percent is reached. Different maturation histories, varying generation and expulsion thresholds all restrict maturity patterns, hydrocarbon resource potential and migration pathways.

Bulk compositional characteristics like API gravity and S/A HC are strongly influenced by thermal maturity and expulsion thresholds. Connate water washing has significant impact on gasoline range aromatic compounds, but little effect on API gravity and S/A HC ratio. Biodegradation affects the gravity of some Family C oils on the western margin of the Madison producing trend. The quality of other oils is not affected by biodegradation. Previous studies attributing Family C oil quality variations to biodegradation could not independently estimate the effects of varying thermal maturity. New correlations between molecular ratios resistant to biodegradation indicate that most gravity differences result from thermal maturity variations. Used judiciously, gravity and gross compositional variations can be employed to focus exploration efforts on targets likely to contain oils of the highest quality.

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OIL FAMILIES IN SOUTHWESTERN SASKATCHEWAN

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ABSTRACT

Three stratigraphically restricted oil families dominate the area south of the Carboniferous subcrop in southwestern Saskatchewan. Oils in the stratigraphic interval of the Mississippian Madison to Lower Cretaceous Mannville of the Battle Creek, Rapdan-Battrum area have > 210°C boiling point, saturate fractions characterized by pristane/phytane ratios < 1, a predominance of even/odd n-alkanes, high C_{23} tricyclic/C_{30} pentacyclic terpanes, a C_{35} hopane prominance, and diasterane/regular sterane ratios <1. These oils are compositionally similar to Family C oils of the Madison subcrop petroleum province in southeastern Saskatchewan and southwestern Manitoba. Work in the eastern Williston Basin has correlated such oils to a Lodgepole source rock. Trisnorhopane thermal maturity ratios (Ts/Tm) suggest that Family C oils in southwestern Saskatchewan are of low thermal maturity (commonly 0.72 < VR < 0.78), and that they were derived from source rocks near their onset of significant hydrocarbon generation (VR = 0.7%). Low thermal maturities result in oils of high density, such that subsequent biodegradation has a more serious affect on oil quality than would be observed if the oils had initially been more mature.

A pervasively biodegraded oil family occurring in the Bakken Formation, particularly near its subcrop edge, has a > 210°C boiling point saturate fraction characterized by pristane/phytane > 1.0 and diasterane/regular sterane > 1.0. In these ways this family of oils resembles Family B oils, an oil family also stratigraphically restricted to the Bakken Formation of southeastern Saskatchewan and southwestern Manitoba. Family B oils in southeastern Saskatchewan are derived from Bakken source rocks. There they have consistently high thermal maturities, Ts/Tm > 1.1, that have been attributed to a delayed expulsion threshold. A noticeable difference between Bakken oils in southwestern Saskatchewan and Family B oils is their consistently low thermal maturity indicators (generally C_{29} sterane S/R < 1.0 and Ts/Tm < 0.6). This suggests either derivation from a source rock of comparable maturity to Lodgepole sources feeding the Rapdan-Dollard Mesozoic oilfield trend or an alteration of thermal maturity indicators by biodegradation.

The third oil family is restricted to Viking reservoirs and is also characterized by pristane/phytane ratios > 1.0 and high diasterane/regular sterane ratios. These oils can be distinguished from oils in the Bakken Formation by their consistently higher percentage of hydrocarbons, high saturate/aromatic hydrocarbon ratios, enrichment in C_{28} steranes and high thermal maturities. Other workers have suggested that the source of these oils is the overlying Upper Cretaceous portion of the Lower Colorado Group shale succession. High thermal maturities result in oils of low density and high quality, such that subsequent biodegradation of these oils, although as pervasive as that affecting other oil families, is not perceived as equally detrimental to oil quality in the Viking.

All three families exhibit variable degrees of biodegradation or water washing. Water washing is inferred by the selective removal of more soluble hydrocarbons, particularly benzene and toluene. Biodegradation is recognized by the selective removal of n-alkanes in both the > 210°C boiling point saturate fraction and the gasoline range. Decreasing heptane value appears to be one of the most systematic variations that accompanies increasing biodegradation. Oil density and API gravity generally vary with heptane value indicating that biodegradation is a dominant control on oil quality in the region. Similar patterns of variation are observed in Bakken oil pools, but the degradation of Viking oils provides one of the strongest correlations between progressive changes in heptane value and API gravity. Biodegraded oils are generally water washed, but water washing is not always accompanied by biodegradation.